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Review of the Proterozoic evolution of the Grenville Province, its Adirondack outlier, and the Mesoproterozoic inliers of the Appalachians

James M. McLelland
Bruce W. Selleck
Department of Geology, Colgate University, Hamilton, New York 13346, USA

M.E. Bickford
Department of Earth Sciences, Syracuse University, Syracuse, New York 13244, USA

ABSTRACT

In recent years, a rapidly expanding database, especially in sensitive high-resolution ion microprobe (SHRIMP) geochronology, has led to significant advances in understanding of the Precambrian tectonic evolution of the Grenville Province, including its Adirondack outlier, and the Mesoproterozoic inliers of the Appalachians. Based upon this information, we review the geochronology and tectonic evolution of these regions and significant similarities and differences between them. Isotopic data, including Pb isotopic mapping, suggest that a complex belt of marginal arcs and orogens existed from Labrador through the Adirondacks, the midcontinent, and into the southwest during the interval ca. 1.8–1.3 Ga. Other data indicate that Mesoproterozoic inliers of the Appalachians, extending from Vermont to at least as far south as the New Jersey Highlands, are, in part, similar in composition and age to rocks in the southwestern Grenville Province. Mesoproterozoic inliers of the Appalachian Blue Ridge likewise contain some lithologies similar to northern terranes but exhibit Nd and Pb isotopic characteristics suggesting non-Laurentian, and perhaps Amazonian, affinities. Models invoking an oblique collision of eastern Laurentia with Amazonia are consistent with paleomagnetic results, and collision is inferred to have begun at ca. 1.2 Ga. The collision resulted in both the ca. 1190–1140 Ma Shawinigan orogeny and the ca. 1090–980 Ma Grenvillian orogeny, which are well represented in the Appalachians. Several investigators have proposed that some Amazonian Mesoproterozoic crust may have been tectonically transferred to Laurentia at ca. 1.2 Ga. Data that potentially support or contradict this model are presented.

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INTRODUCTION

A schematic map of the surface and subsurface extent of the Grenville orogen is shown in Figure 1, including the Grenville Province, as well as the ca. 1.2–0.98 Ga Mesoproterozoic inliers of the Appalachians (Rankin et al., 1993, their Plate 5B). Additional exposures of the Grenville orogen exist in the Mesoproterozoic inliers of Texas, lying to the west in Figure 1 (Rankin et al., 1993). Rivers et al. (1989) divided the exposed Grenville Province into three principal tectonic segments (Fig. 1): (1) the Parautochthonous Belt, (2) the Allochthonous Polycyclic Belt, and (3) the Allochthonous Monocyclic Belt. The Parautochthonous Belt consists of reworked Archean and Paleoproterozoic rocks of the Superior Province foreland that were deformed by northwestward thrusting of the multiply metamorphosed ca. 1.7–0.98 Ga Allochthonous Polycyclic Belt along the Allochthon Boundary thrust at ca. 1050 Ma, followed by orogen collapse at ca. 1020 Ma. A final contractional pulse (ca. 1010–980 Ma) ending with extensional collapse gave rise to the Grenville Front thrust. The Allochthonous Monocyclic Belt consists of rocks younger than ca. 1.35 Ga. Details of these belts and the orogenic sequence of events preserved in them will be discussed in later sections. The subsurface distribution of the Grenville orogen in Figure 1 has been approximated from drill-core samples and geophysical data (Rankin et al., 1993) and is especially uncertain below the complexities of the northeastern Appalachian orogen. However, a small outcropping of Mesoproterozoic basement at Cape Breton (Fig. 1) suggests the existence of Mesoproterozoic basement to the west.

Figure 1 underscores the continental scale of the Grenville orogen and inevitably leads to questions regarding relationships

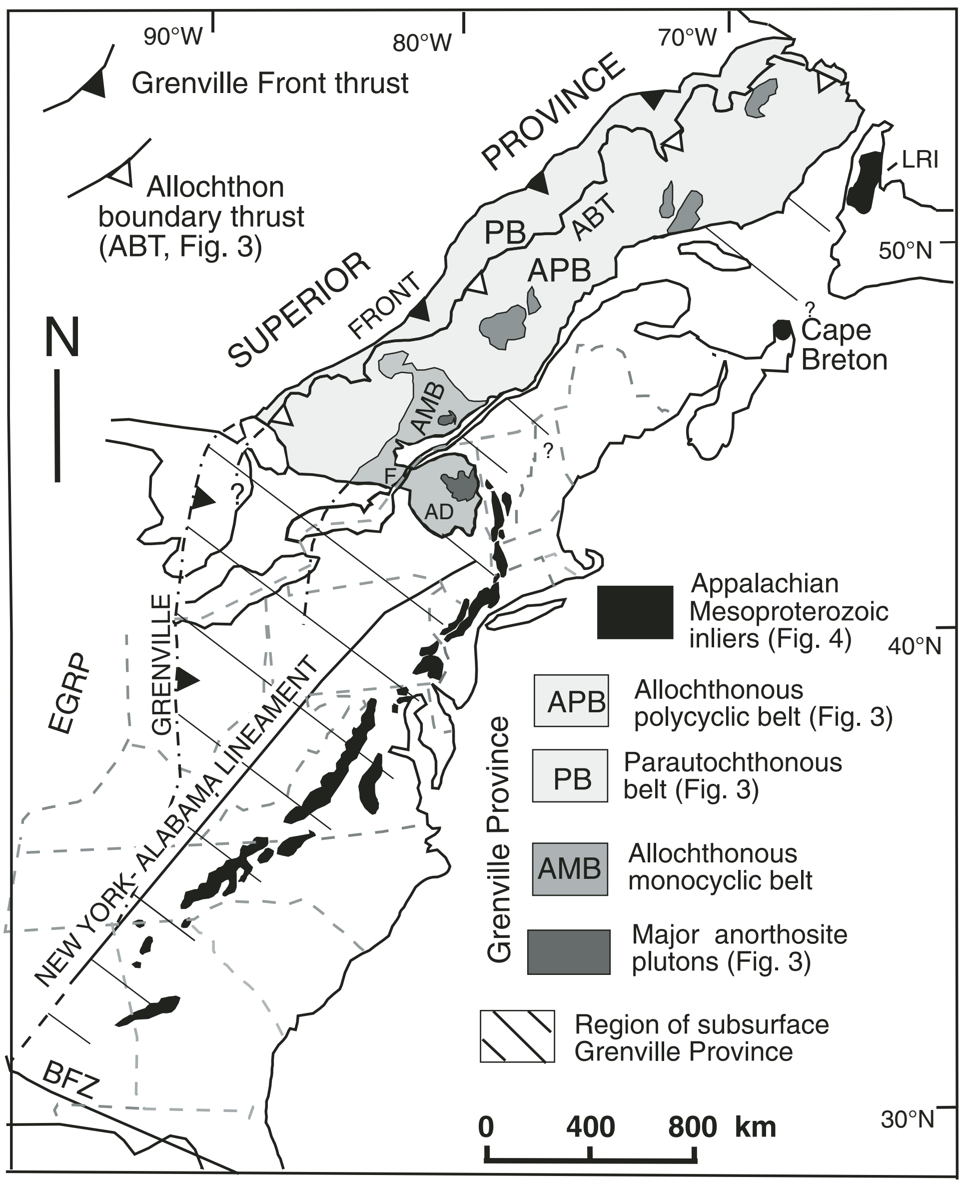

Figure 1. The extent of the Grenville orogen in eastern North America. The Grenville Province, its tectonic divisions, and major anorthosite massifs are shown in shades of gray identified in the legend. Mesoproterozoic inliers of the Appalachians are designated in black. The subsurface distribution of Grenville basement as determined from drill-hole and geophysical data is represented by diagonal ruling. The New York–Alabama Lineament (King and Zeitz, 1978) marks the trend of a large magnetic anomaly that may have tectonic implications. BFZ—Bahamas fracture zone; EGRP—Eastern granite-rhyolite province. (Figure was modified after Rivers et al., 1989; Tollo et al., 2004.)

between its various components and their plate-tectonic evolution. In recent years, an expanding database derived from modern geophysics, petrology, tectonics, numerical modeling, field investigations, and geochronology has led to substantial clarification of these issues. Most geochronological studies performed prior to the last decade involved variations of thermal ionization mass spectrometry (TIMS), which offers greater analytical precision but less spatial resolution than sensitive high-resolution ion microprobe (SHRIMP) techniques. As a result, with TIMS, it is much more difficult to date and analyze the small and intricately intergrown domains and zones that characterize zircons in polymetamorphic rocks. We suggest that some current disparities in ages cited in the literature are the result of comparisons between SHRIMP- and TIMS-derived data. Based on experience and reading of pertinent geologic literature, we have found that SHRIMP-derived ages are generally more reliable in deciphering complex metamorphic and igneous histories. Our results indicate that in the Adirondacks, as well as in the southern Appalachians (e.g., Tollo et al., 2006; Southworth et al., this volume; Volkert et al., this volume), granitoids originally dated by TIMS techniques, and subsequently re-dated by SHRIMP, commonly yield ages that differ by 10–50 m.y. Table DR1 (in the GSA Data Repository[1]) presents a summary of both SHRIMP and TIMS U-Pb geochronological data for the Adirondacks, and similar data for the Appalachians can be found in other papers in this volume (e.g., Southworth et al., Tollo et al., Volkert et al., this volume). Rivers (1997, 2008) has summarized geochronological data for the entire Grenville Province, and Figure 2, which is consistent with his orogenic definitions, provides a time scale for events in the Grenville orogen and illustrates some of the important geochronological links that exist across much of the region. In addition to U-Pb zircon igneous crystallization ages for plutons, whole-rock Nd model ages (e.g., Dickin, 2000; Daly and McLelland, 1991) provide estimates for the time of extraction of magma from the mantle and the extent to which plutons intruding the crust may be juvenile. Zircon, titanite, and monazite (e.g., Mezger et al., 1991, 1992; Heumann et al., 2006) ages have been used to constrain the timing of metamorphic events and, together with $^{40}Ar/^{39}Ar$ ages, have facilitated the determination of cooling histories (e.g., Cosca et al., 1991; Busch et al., 1996; Dahl et al., 2004; Streepey et al., 2000, 2001; Rivers, 2008). Equally significant data are Pb-Pb whole-rock isotopic ages that serve to establish the age and nature of crustal Pb reservoirs (Sinha et al., 1986; Sinha and McLelland, 1999; Loewy et al., 2003). All of these will be discussed in the main body of the text and will be utilized to provide a more unified perception of the Grenville orogen. In the following, we summarize the current state of understanding in the regions shown in Figure 1, seek the similarities and differences between them, and examine models that may help to account for their plate-tectonic evolution.

[1]GSA Data Repository Item 2010XXX, **[[***Need short description***]]**, is available at www.geosociety.org/pubs/ft2010.htm, or on request from editing@geosociety.org, Documents Secretary, GSA, P.O. Box 9140, Boulder, CO 80301-9140, USA.

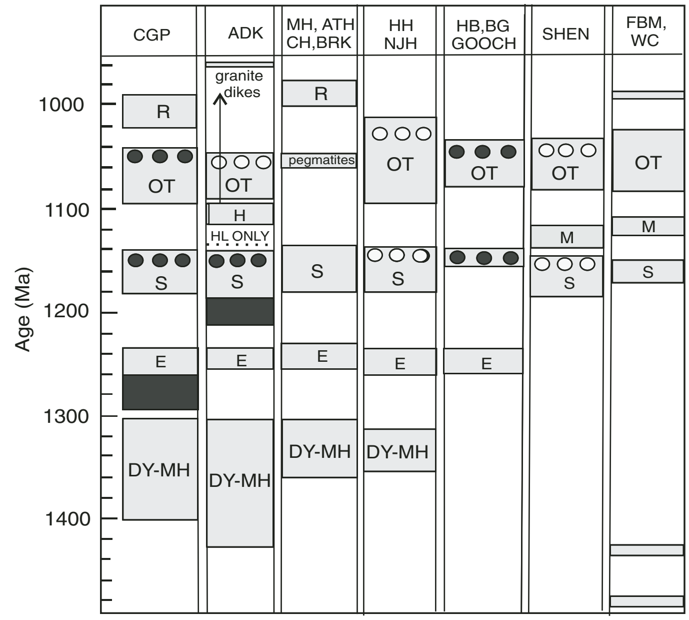

Figure 2. Time scale for Mesoproterozoic events in eastern North America. Orogenies: R—Rigolet, OT—Ottawan, S—Shawinigan, and E—Elzevirian. Symbols: Black dots—anorthosite-mangerite-charnockite-granite magmatism; white dots— mangerite-charnockite-granite magmatism; black—precollision calc-alkaline magmatism. Abbreviations: CGP—Grenville Province of Canada (Rivers, 2008); ADK—Adirondack outlier (McLelland, this volume). References for geochronology in the following abbreviated localities may be found in the text (under Mesoproterozoic Inliers of the Appalachians): AH—Adirondack Highlands; ATH—Athens Dome, CH—Cheshire Dome, BRK—Berkshire Mountains; DY-MH—Dysart–Mount Holly suite; FBM—French Broad massif; HH—Hudson Highlands; M—Mars Hill; MH—Mount Holly complex; NJH—New Jersey Highlands; HB—Honey Brook uplands; BG—Baltimore Gneiss Domes; Gooch—Goochland terrane; H—Hawkeye granite; HL—Highlands; SHEN—Shenandoah massif; M—Marshall granite; TR—Trimont Ridge; WC—Wolf Creek.

GEOLOGICAL OVERVIEW OF THE GRENVILLE PROVINCE AND ITS ADIRONDACK OUTLIER

Introduction

The Grenville Province (Figs. 1 and 3) consists primarily of reworked amphibolite- to granulite-facies crust ranging in age from Archean to ca. 980 Ma and characterized by province-wide occurrence of ca. 1000–970 Ma K-Ar and ^{40}Ar-^{39}Ar metamorphic mineral ages (Stockwell, 1964; Davidson, 1998). The Grenville Front (Figs. 1 and 3) is a crustal-scale, ca. 1000 Ma southeast-dipping reverse fault that lies along the northwestern margin of the province, where it separates low-grade Archean and Paleoproterozoic rocks of the foreland from the tectonically reworked parautochthon, within which foreland basement has been variably displaced to the northwest via progressive understacking. Members of the ca. 1237 Ma Sudbury dike swarm (Fig. 3) cross the Grenville Front and are common in

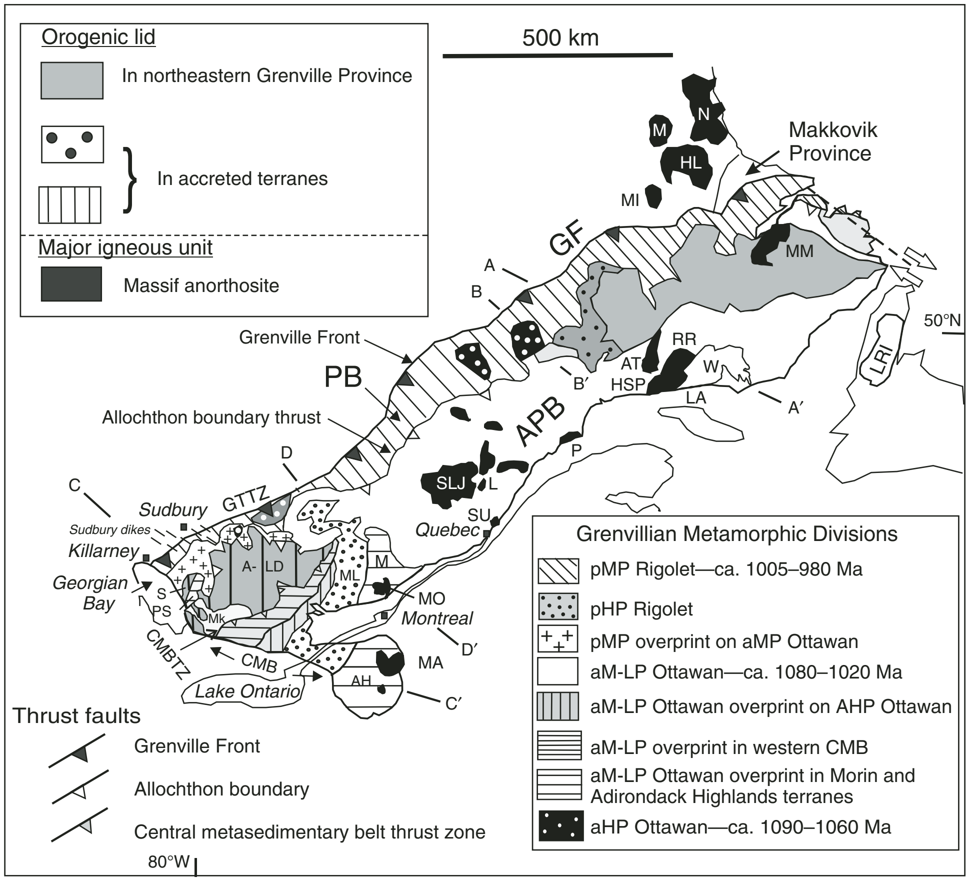

Figure 3. Generalized tectonic divisions of the Grenville Province. Cross-section lines for Figure 8 are shown. APB—Allochthonous Polycyclic Belt; PB—Parautochthonous Belt; A-LD—Algonquin and Lac Dumoine domain; AH—Adirondack Highlands; CMB—Central metasedimentary belt; CMBTZ—Central metasedimentary belt thrust zone; GFTZ—Grenville Front tectonic zone; ML—Mont Laurier domain; M—Morin terrane; PS—Parry Sound domain; S—Shawanaga domain; W—Wakeham basin. Major anorthosite massifs: MA—Marcy (ca. 1150 Ma); MM—Mealy Mountains; MO—Morin; LA—Lac Allard; LSJ—Lac St. Jean; HSP—Havre-St. Pierre; A—Atikonak; HL—Harp Lake; L—Labrieville; M—Michikamau; MI—Mistastin; N—Nain-Kiglapait; P—Pentecote; RR—Riviere Romaine; SU—St. Urbain. Cities: O—Ottawa; Q—Quebec City; S—Shawinigan Falls. Pressure defined subdivisions: pMP–parautochthonous medium pressure belt; aM-LP–allochthonous medium- to low-pressure belt; aHP–allochthonous high-pressure belt; pHP–parautochthonous high-pressure belt. (Figure was modified after Rivers, 2008.)

the parautochthon (Bethune and Davidson, 1997). At the eastern margin of the Grenville Province, the Grenville Front swings southeast (Fig. 3), transitioning to a dextral lateral-ramp strike-slip fault (Gower et al., 2008). At map scale, the Grenville Front appears continuous, rectilinear, and is almost 2000 km in length; however, in reality, it is neither a single fault nor single mylonite zone, although continuous mylonitic segments do exist over large distances (Davidson, 1998).

The Allochthon Boundary thrust (Figs. 1 and 3) is a high-grade, gently dipping thrust fault that lies southeast of the Grenville Front and structurally above the Parautochthonous Belt. It possesses ramp-flat geometry, soles out in the middle to lower crust, and overlies the parautochthon from Georgian Bay to eastern Labrador (Rivers et al., 1989). The lobate map-scale outcrop pattern of the thrust (Fig. 3) attests to its shallow dip (Davidson, 1998). Seismic profiles indicate that, beneath the thrust sheets, Paleoproterozoic and Archean rocks of the parautochthon continue to the south for up to 150 km (Ludden and Hynes, 2000). Rocks within the allochthons do not directly correlate with units in the foreland and are interpreted as far-traveled portions of southeastern Laurentia. The Sudbury dike swarm (Fig. 3) does not cross the Allochthon Boundary thrust, consistent with the inferred major displacement of the latter (Bethune and Davidson, 1997). The ca. 1.7–1.4 Ga medium- and high-grade metamorphic assemblages in the Allochthonous Belt are commonly overprinted by 1080–980 Ma Grenvillian metamorphism. In this sense, the allochthon has experienced several orogenic pulses and is referred to as the Allochthonous Polycyclic Belt (Figs. 1 and 3). Thick mylonites and high-pressure (i.e., intact or retrograded eclogite) assemblages are common at the base of the thrust sheets and record high-pressure belts (Fig. 3) formed during Grenvillian telescoping of the continental margin toward the northwest (Rivers, 2008).

The ca. 1.35–1.02 Ga Adirondack Mountains of New York State (Fig. 4) provide a bridge between the large (2000 × 600 km) Grenville Province (ca. 1700–980 Ma), of which it is an outlier, and the (ca. 1.35–0.98 Ga) inliers of the Appalachian Mountains of the eastern United States (Fig. 1). The Adirondacks form a slightly elliptical dome divisible into the Adirondack Highlands terrane, consisting mostly of orthogneiss, and the smaller Adirondack Lowlands domain (Fig. 4), underlain mainly by marble-rich metasediments of shallow-marine origin. The Adirondack Lowlands connect to the Frontenac domain (Fig. 5) across the southern reaches of the St. Lawrence River. Carr et al. (2000) proposed the term Frontenac–Adirondack Belt, noting that these two areas share 1190–1140 Ma plutonism and metamorphism related to the Shawinigan orogeny (Fig. 2). An alternative reconstruction utilized here, and detailed later, couples the Adirondack Lowlands

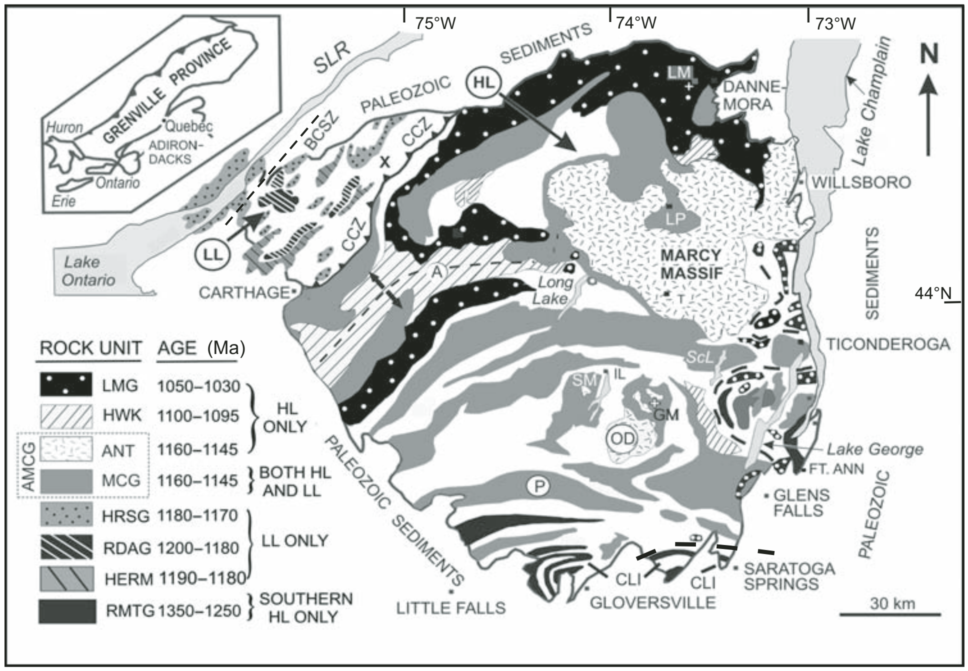

Figure 4. Generalized geological map of the Adirondacks. Units designated by patterns and initials consist of metaplutonic rocks dated by U-Pb zircon geochronology, with group ages indicated in legend. Units present only in the Adirondack Highlands (HL) are: RMTG—Royal Mountain tonalite and granodiorite (southern HL only), HWK—Hawkeye granite, LMG—Lyon Mountain Granite, and ANT—anorthosite. Units present in the Adirondack Lowlands (LL) only are: HSRG—Hyde School and Rockport granites (Hyde School also contains tonalite); RDAG—Rossie diorite and Antwerp granodiorites; HERM—Hermon granite. Granitoid members of the anorthosite-mangerite-charnockite-granite suite (AMCG) are present in both the Adirondack Highlands and Lowlands. Unpatterned areas consist of metasediments, glacial cover, or undivided units. A—Arab Mountain anticline; BCSZ—Black Creek shear zone; CLI—Canada Lake isocline; CCZ—Carthage-Colton shear zone; IL—Indian Lake; LM—Lyon Mountain; LP—Lake Placid; OD—Oregon Dome; P—Piseco antiform; SM—Snowy Mountain Dome; T—Tahawus. Dashed line shows approximate western limit of migmatitic metapelites in the eastern Adirondacks. (Figure was modified after McLelland et al., 2004.)

and Frontenac domain with ca. 1.3–1.2 Ga accreted terranes to the west, thus corresponding to Wynne-Edwards (1978) original Central metasedimentary belt (Fig. 5, CMB). We also introduce a separate, more easterly, belt consisting of the Adirondack Highlands and the Mount Holly complex, Vermont (Fig. 5), referred to as the Adirondack Highlands–Mount Holly Belt. The bedrock Proterozoic geology of this belt is similar throughout with quartzites, marbles, metapelites, and many plutonic rocks of the Adirondacks, extending along strike into the Mount Holly complex (Ratcliffe et al., 1991, p. 78, 80, 90; Karabinos et al., 1999, 2003) and providing an almost direct connection of the Adirondacks to the Mesoproterozoic inliers of the Appalachians.

Early Geological Evolution of the Grenville Province (2.0–1.3 Ga)

At its far western and eastern ends, the Grenville Front truncates ca.1.9–1.8 Ga rocks (Figs. 3 and 5), an age interval that coincides with the Penokean and Makkovik orogenies (Dickin and McNutt, 1989). These observations suggest that the southeastern margin of Paleoproterozoic Laurentia may have hosted an Andean-type arc developed in a combined Penokean-Makkovik-Ketilidian orogenic belt (Hoffman, 1989). In the parautochthon south of Sudbury, Dickin and McNutt (1989) described a major belt of calc-alkaline granitic gneiss with Nd model ages of 2.09–1.72 Ga (Barillia, Algonquia, Fig. 6). The belt, which is 50–150 km wide, extends southeastward for several hundred kilometers through the Algonquin–Lac Dumoine terrane (Fig. 6) and has been interpreted as an accreted Penokean arc (Dickin and McNutt, 1989; Dickin et al., 1990; Dickin, 2000). This interpretation is consistent with the age interval and nearby ductile deformation dated at ca. 1.85 Ga (Zolnai et al., 1975). However, the 1.9–1.72 Ga Nd model ages could also be due to mixing of Archean crust with Mesoproterozoic magmas, as demonstrated by DeWolf and Mezger (1994) for other rocks closer to the Grenville Front. The issue remains unresolved.

An extensive record of Paleoproterozoic arcs along the Laurentian margin is preserved northwest of Sudbury (Fig. 3), where ca. 1.7 Ga granitoids within the Killarney magmatic belt are inferred to represent continuations of the Yavapai (1.8–1.7 Ga) and Mazatzal (1.7–1.6 Ga) accretionary orogens of the southwestern United States (van Breemen and Davidson, 1988; Davidson, 1998; Karlstrom et al., 2001; Rivers and Corrigan, 2000). They sit within the Southern Province of the Canadian Shield and within the Grenville Front tectonic zone (Fig. 3), where they are deformed. Gower and Krogh (2002) summarized events at the eastern end of the Grenville Province in Labrador, where ca. 1680–1655 Ma accretion of outboard arcs as old as 1.71 Ga resulted in plutonism and metamorphism (1665–1655 Ma). This was followed by intrusion of the ~800-km-long, juvenile, ca. 1654–1646 Ma Trans-Labrador Batholith, which represents a calc-alkaline, Andean-style arc (Gower and Krogh, 2002). During the waning stages of this magmatism, emplacement of a trimodal suite of mafic-anorthositic-monzogranitic magmatism ensued until ca. 1.61 Ga, and the ca. 1.63 Ga Mealy Mountains anorthosite was emplaced. Gower and Krogh (2002) refer to these 1.71–1.6 Ga events as the

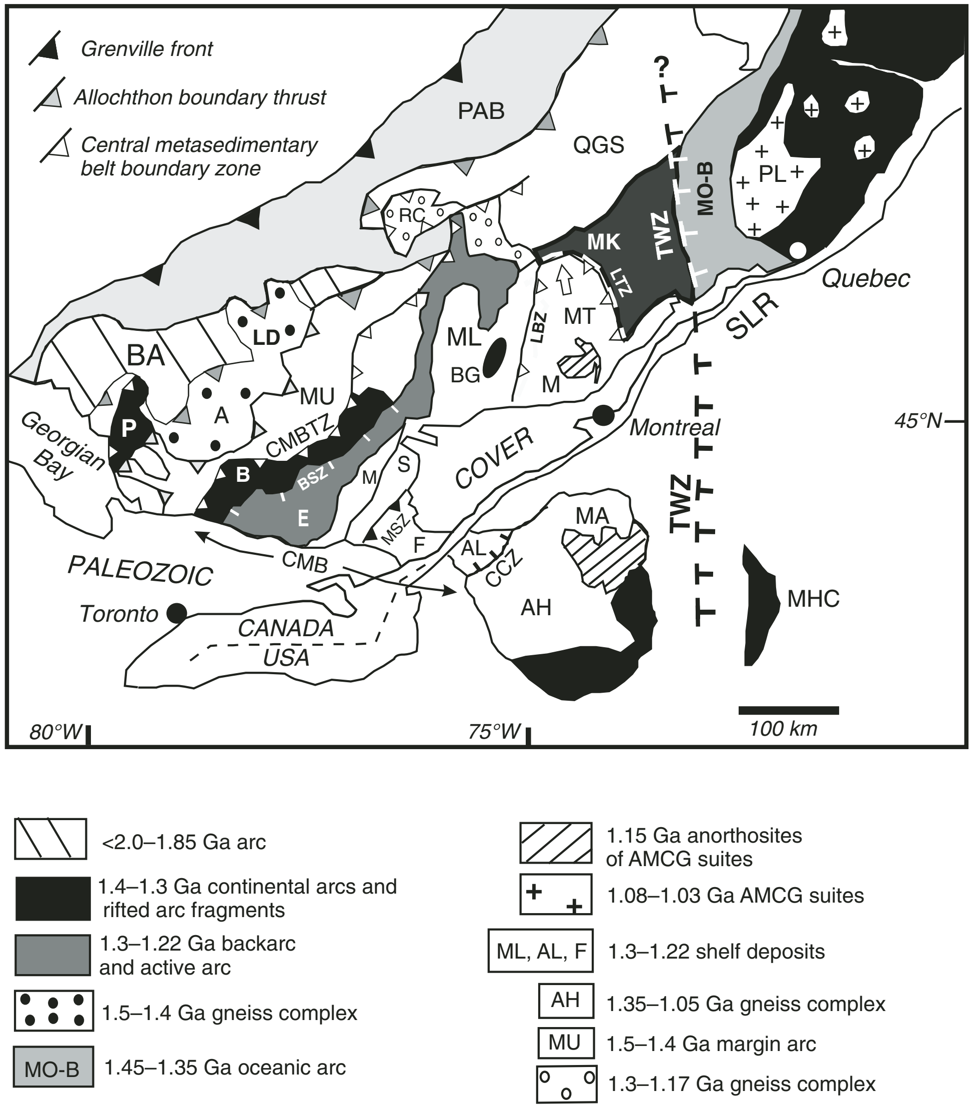

Figure 5. Generalized geologic map of the southwestern segment of the Grenville Province plus the Mount Holly complex. Vermont. A, LD—Algonquin–Lac Dumoine domain; AH—Adirondack Highlands; AL—Adirondack Lowlands; B—Bancroft terrane; BA—Barillia (Dickin, 2000); BSZ—Bancroft shear zone; BG—Bondy Gneiss Dome; CCZ—Carthage-Colton shear zone; CMB—Central metasedimentary belt; CMBTZ—Central metasedimentary belt thrust zone; E—Elzevir terrane; F—Frontenac terrane; LBZ—Labele shear zone; LTZ—Lac Taureau shear zone; M—Mazinaw terrane; MHC—Mount Holly complex; MK—Mekanic terrane; ML—Mont Laurier terrane; MM—Marcy anorthosite; MO—Morin anorthosite; MO-B—Montauban–La Bostonaiss arc; MU—Muskoka domain; MSZ—Maberly shear zone; P—Parry Sound domain; PAB—Parautochthonous Belt; PL—Parc des Laurentides anorthosite-mangerite-charnockite-granite suite; QGS—Quebec gneiss segment; RC—Reservoir Cabonga terrane; SLR—St. Lawrence River; TWZ—Tawachiche shear zone. (Figure was modified after Hanmer et al., 2000.)

Labradorian orogeny. The juvenile nature of the crust is manifested by Nd model ages of ca.1.71 Ga.

In Labradoria (Fig. 6), as defined by Dickin (2000) on the basis of Nd model ages, ca. 1.7–1.6 Ga crust was metamorphosed and intruded by 1.52–1.46 Ga continental margin arc plutons during the Pinwarian orogeny (Gower and Krogh, 2002). Plutons of ca. 1.5–1.4 Ga age occur throughout much of the Grenville Province (Ketchum et al., 2004) and are common in the Algonquin–Lac Dumoine and Britt terranes near Georgian Bay in southwest Ontario (Fig. 6) and continue through Ontario and Quebec, where large tracts characterized by ca. 1.55 Ga Nd model ages imply juvenile crust (Dickin and McNutt, 1989; Dickin and Higgins, 1992; Dickin, 2000). These results suggest the presence of a southeast-facing, long-lived arc along the Laurentian margin (Rivers and Corrigan, 2000; Karlstrom et al., 2001; Rivers, 2008), which is informally referred to as the Britt-Pinware granitoid belt and has been suggested to be continuous with the 1.5–1.4 Ga Eastern granite-rhyolite province of the midcontinent, thus implying an ~4000-km-long, northwest-subducting Andean-type arc (with back-arc basins) from Labrador to New Mexico.

From ca. 1.45 to 1.3 Ga, Andean-style magmatism shifted southeastward in the Grenville Province and was active from the Mekinac terrane (Fig. 5) to the current location of the Central metasedimentary belt thrust zone (CMBTZ, Figs. 3 and 5), with northwest polarity throughout. Outboard of Laurentia, the Montauban arc (Figs. 3 and 5) developed as early as 1.44 Ga and accreted to Laurentia at ca. 1.39 Ga (Corrigan and van Breemen, 1996). Accretion took place approximately along the Tawachiche shear zone (TSZ, Fig. 5), which was later reactivated as a mid- to late-Ottawan detachment fault (Corrigan and van Breemen, 1996). At ca. 1.39 Ga, the arc was thrust over ca. 1.45–1.37 Ga tonalitic gneisses of the Mekinac terrane (Fig. 5), and intruded by ca. 1.39–1.37 Ga tonalitic magmas of the La Bostonnais suite, which intruded the Montauban arc. Within the Mekinac terrane,

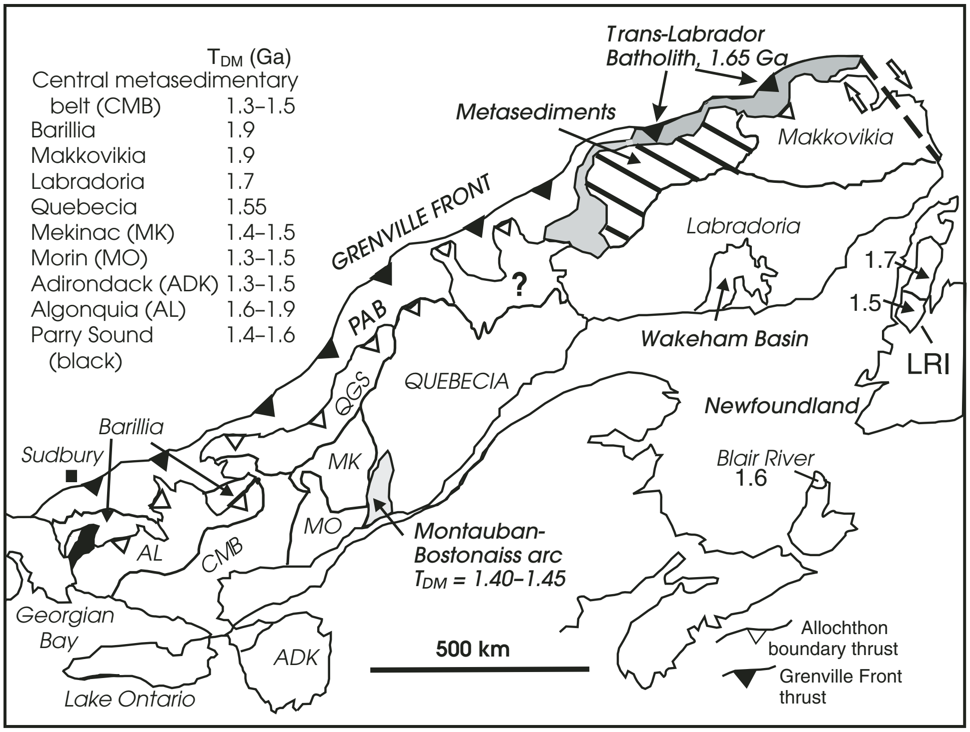

Figure 6. Subdivisions of the Grenville Province based on Nd-model ages. ADK—Adirondack Mountains; AL—Algonquia (includes Lac Dumoine); CMB—Central metasedimentary belt; LRI—Long Range inlier; MK—Mekinac terrane; MO—Morin terrane; PAB—parautochthonous belt; QGS—Quebec Gneiss segment. Arrows represent dextral offset (Gower, 2008) along lateral ramp fault. (Figure was modified from Dickin, 2000.)

tonalitic orthogneiss skirts the north end of the Morin terrane and forms the structural basement for the Mount Laurier domain of the Central metasedimentary belt, where ca. 1.39 Ga orthogneiss is exposed in gneiss domes, e.g., Bondy and LaCoste domical complexes (Corriveau and van Breemen, 2000; Wodicka et al., 2004). Farther south, tectonic slices of ca. 1.4–1.3 Ga tonalitic to granitic orthogneiss (Dysart suite) within the Central metasedimentary belt thrust zone (CMBTZ, Figs. 3 and 5) are similar to, and may be continuous with, the 1.45–1.4 Ga Eastern granite-rhyolite province (Rivers and Corrigan, 2000; Slagsted et al., 2009). This entire belt contains Nd model ages of 1.55–1.4 Ga, which, together with the zircon crystallization ages, attest to the juvenile nature of the continental margin arc.

Ca. 1.3 Ga Rifting of the Ca. 1.4–1.3 Ga Continental Margin Arc, Opening of the Central Metasedimentary Belt Backarc Basin, and the Elzevirian Orogeny (ca. 1245–1220 Ma)

McEachern and van Breemen (1993) and Hanmer et al. (2000) proposed that the 1400–1300 Ma Andean margin in the region of the present-day Bancroft zone (Fig. 7A) rifted along a northeast-trending fault zone at ca. 1.3 Ga. The rift split the arc, leaving the western remnants behind (Fig. 7A) as the Dysart suite, whereas the eastern remnants served as the foundations for island arcs in the Central metasedimentary (Dickin and McNutt, 2007) and Adirondack–Mount Holly belts (Fig. 7A) and are referred to as the Dysart–Mount Holly suite. The thick Irving Pond orthoquartzite of the southern Adirondacks is spatially associated with ca. 1.4–1.3 Ga Dysart–Mount Holly fragments and contains detrital zircons ranging in age from ca. 2.5 to 1.31 Ga. We interpret the quartzite as having been deposited on the Laurentian mainland, in proximity to the 1.4–1.3 Ga arc prior to 1.3 Ga, and subsequently transported eastward on rifted arc fragments.

Emplacement of ca. 1.29–1.25 Ga nepheline syenite (Lumbers et al., 1990), as well as gabbro (Pehrsson et al., 1996), accompanied rifting along the western margin of the evolving basin (Fig. 7A), and primitive tholeiitic arcs formed outboard (Heaman et al., 1987; Davis and Bartlett, 1988). From ca. 1.3 to 1.22 Ga, the entire region evolved as a backarc basin (Harnois and Moore, 1991; Carr et al., 2000) that bordered the eastern margin of Laurentia and accumulated a thick section of sediments (Fig. 7B). Marble and volcanics with backarc geochemical characteristics (Holm et al., 1986) are important constituents of the lithic package, except within the Frontenac terrane, where marble is subordinate to quartzitic and pelitic metasediments, and volcanic rocks are lacking. The oldest rocks in the basin are ca. 1296 Ma bimodal rhyolites and mafic volcanics (Silver and Lumbers, 1965; Davis and Bartlett, 1988) and the >1279 Ma Tudor Formation of the Mazinaw domain, which may contain ophiolitic material (Harnois and Moore, 1991; Corfu and Easton, 1995, 1997). Additional ophiolitic fragments occur in the Elzevir terrane (Easton and Ford, 1994; Smith and Harris, 1996), and a large occurrence of ophiolite has been reported in the Adirondack Lowlands (Chiarenzelli et al., 2007). Within the backarc basin, deformation was followed by intrusion of ca. 1276–1270 Ma calc-alkaline plutons in the Mazinaw terrane (Corfu and Easton, 1995), and the ca. 1267 Elzevir tonalitic batholith (Easton and Ford, 1994)

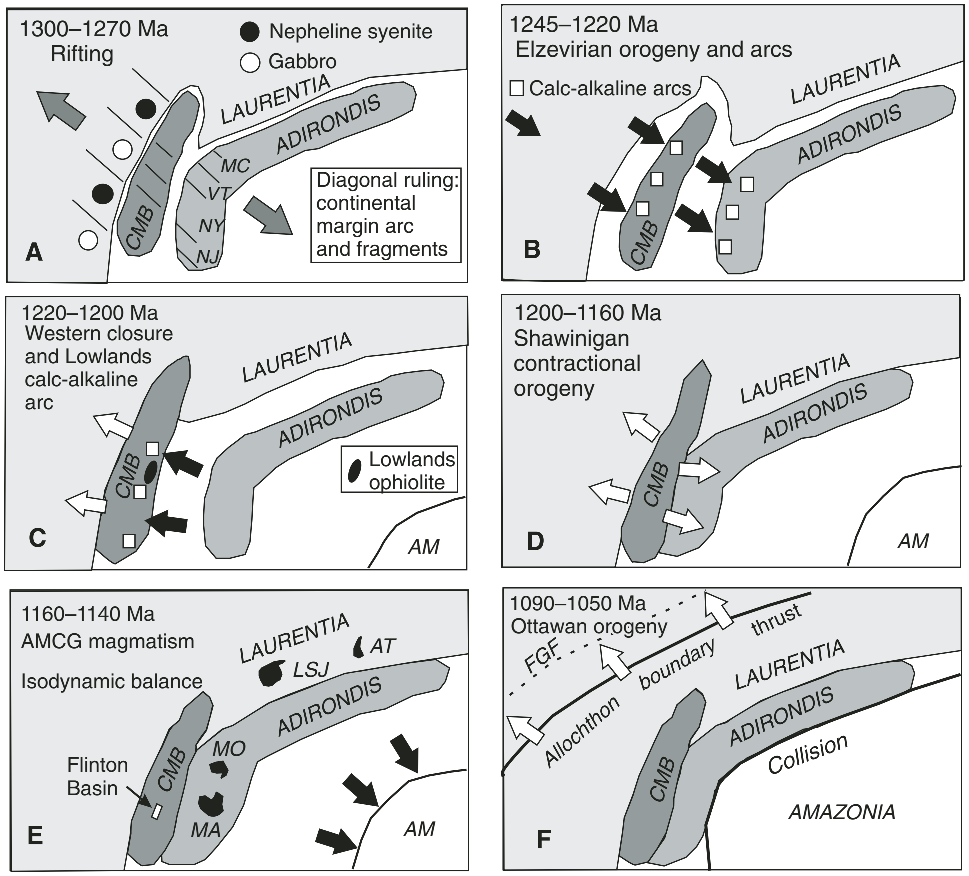

Figure 7. Schematic panel diagram summarizing the distribution and interaction of various segments of northeastern North America during the interval ca. 1300–1050 Ma, encompassing the rifting, opening, and closing of the Central metasedimentary belt (CMB) and the Ottawan collision with Amazonia. Blocky black arrows indicate the polarity of subduction. Blocky open arrows indicate direction of thrusting. Blocky gray arrows represent extension. The model is derived, in part, from Gower (1996), who proposed the term Adirondis to include crust that was rifted from Laurentia at ca. 1.4–1.3 Ga and reattached by collision ca. 1.2–1.0 Ga. As defined here, Adirondis also includes crust of Mauricie (Quebec), Adirondack, and northern Appalachian Mesoproterozoic crust to as far south as the New Jersey Highlands. Also included is Grenvillian-age crust currently buried beneath Paleozoic strata south of the St. Lawrence River. See text for details. Note that Amazonia is inferred to have remained south of the Adirondis region until ca. 1090 Ma, but its regional plate motion is thought to have helped promote closure of the Central metasedimentary belt rift and the Shawinigan collision. Abbreviations: AM—Amazonia; AT—Atikonak River anorthosite; FSG—future Grenville Front; LSJ—Lac St. Jean anorthosite; MA—Marcy anorthosite; MC—Mauricie area, Quebec; MO—Morin anorthosite; NJ—New Jersey; NY—New York; VT—Vermont.

above an eastward-directed subduction zone (Fig. 7B). Bimodal backarc magmatism occurred from ca. 1.26 to 1.25 Ga. From ca. 1245 to 1220 Ma, gabbroic, granodioritic, and granitic stitching plutons intruded across the entire belt and were accompanied by metamorphism (Elzevirian orogeny) associated with arc amalgamation (Carr et al., 2000). A spectrum of similar ca. 1.27–1.22 Ga rocks is present in the Adirondack–Mount Holly belt (McLelland and Chiarenzelli, 1990; Ratcliffe et al., 1991). Within the Adirondacks, metapelitic migmatites contain detrital zircons ranging in age from ca. 1.32 to 1.22 Ga (Heumann et al., 2006). The absence of Archean zircons suggests deposition in a restricted basin outboard of Laurentia. Bickford et al. (2008) interpreted the protoliths to have been graywacke-shale sequences associated with postrifting arc magmatism.

Ratcliffe and Aleinikoff (2008) and Volkert et al. (this volume) demonstrated that the Dysart Mount Holly suite extends through the Hudson Highlands and at least as far south as the New Jersey Highlands. On the western side of the rift, Dickin and McNutt (2007) used geophysical data to extend its margin southwest to similar latitude. These results indicate that the basin was large, and Rivers and Corrigan (2000) estimated its size to have been on the order of the Sea of Japan. We further extend the eastern margin of the rift by combining it with Gower's (1996) Adirondis block (Fig. 7A), which rifted from Laurentia at ca. 1.4–1.3 Ga. In this paper, we refer to the combined terrane as Adirondis.

Cessation of arc magmatism at ca. 1.22 Ga is inferred to have coincided with the early stages of closure of the western backarc basin (Corrigan and Hanmer, 1995; Corriveau and van Breemen, 2000; Wodicka et al., 2000, 2004), followed by ca. 1.19 Ga northwest-directed thrusting (McEachern and van Breemen, 1993; Hanmer and McEachern, 1992) over the Muskoka domain (Fig. 7C). This western closure was accompanied by stepping out of subduction to the east and a change in polarity to the northwest, resulting in calc-alkaline magmatism in the Adirondack Lowlands (Wasteneys et al., 1999), which, at that time, formed the leading edge of Laurentia (Fig. 7C). Deformation extending from ca. 1.19 to 1.17 Ga within of the allochthonous monocyclic belt (Hanmer and McEachern, 1992; McEachern and van Breemen, 1993; Nadeau and van Breemen, 1998; Corriveau and van Breemen, 2000; and Wodicka et al., 2004) is interpreted to reflect the accretion of Adirondis to Laurentia and the onset of the Shawinigan orogeny. Prior to closure, most basin rocks were metamorphosed at amphibolite-facies grade by ca. 1245–1220 Ma contractional pulses of the Elzevirian orogeny related to arc amalgamation.

Allochthonous Monocyclic Belt and the Shawinigan Orogeny (Ca. 1190–1140 Ga)

The allochthonous monocyclic belt consists of the Central metasedimentary belt, the Adirondack Highlands, and the Morin terranes (Figs. 3 and 6), and preserves only ages younger than ca. 1.3 Ga. The large, southeast-dipping Central metasedimentary boundary thrust zone (CMBTZ, Figs. 3 and 6) borders the belt on the northwest and exhibits thrusting to the northwest that occurred on at least two occasions (i.e., 1190–1174 Ma and 1080 to1050 Ma; Carr et al., 2000). The earlier of these events reflects the Shawinigan orogeny (Fig. 2), which was defined by Rivers and Corrigan (2000) as occurring in the interval ca. 1190–1140 Ma. Within the Canadian Grenville Province, effects of the Shawinigan orogeny have been recognized only in the Frontenac, Mont Laurier, Cabonga, and Morin terranes, where ca.1180–1160 Ma amphibolite-facies assemblages and plutons are preserved (Marcantonio et al., 1990; Martignole, 1996; Wodicka et al., 2004). In contrast, a broad spectrum of Shawinigan plutonism and metamorphism is exceptionally well developed in the Adirondacks and appears to be widespread in the Appalachian inliers. Accordingly, a relatively detailed account of Shawinigan features in the Adirondacks is presented here.

The type locality for the Shawinigan orogeny (Rivers, 1997; Rivers and Corrigan, 2000) is near the town of that name in the Mauricie region of Quebec, where Corrigan (1995) recorded evidence of 1190–1160 Ma tectonism. Subsequently, the orogenic interval was expanded to 1190–1140 Ma (Rivers, 1997; Rivers and Corrigan, 2000) in order to include TIMS ages of metamorphic zircon and mylonitization in the Cabonga, Mont Laurier, and Morin terranes (Friedman and Martignole, 1995; Martignole, 1996; Martignole and Friedman, 1998). However, these authors emphasize that high-grade metamorphism and associated tectonism are largely restricted to the interval 1190–1160 Ma and preceded emplacement of the 1155 ± 3 Ma Morin anorthosite-mangerite-charnockite-granite suite (Doig, 1991). In the Mont Laurier terrane, U-Pb zircon SHRIMP dating of the monzonite-gabbro Chevreuil suite (ca. 1170–1160 Ma) demonstrates that the suite postdates regional ca. 1200–1180 Ma granulite-facies metamorphism and deformation of the ca. 1390 Ma Bondy complex (Wodicka et al., 2004). Zircon TIMS analysis of an undeformed pegmatite in a large shear zone yields a crystallization age of 1156 ± 2 Ma and is overprinted by amphibolite-grade metamorphism (Corriveau and van Breemen, 2000). These results document high-grade metamorphism and tectonism at ca. 1200–1160 Ma, followed by anorthosite-mangerite-charnockite-granite intrusion that prolonged elevated temperatures on a regional scale. Renewed contractional pulses may have occurred locally, but widespread tectonism had terminated by ca. 1160 Ma. Similar results have been reported from the Frontenac domain (Davidson, 1998), where undeformed, unmetamorphosed members of the ca. 1160 Ma Kingston dike swarm crosscut 1176–1162 Ma deformed plutons and ca. 1180–1160 Ma granulite-facies assemblages (Marcantonio et al., 1990). Within the Adirondack Lowlands, titanite and monazite preserve ages as old as 1156 Ma and 1171 Ma, respectively (McLelland et al., 1988a; Mezger et al., 1991, 1992; Heumann et al., 2006), and no post–1172 Ma regional tectonism or metamorphism is present. Local anorthosite-mangerite-charnockite-granite plutons were emplaced at ca. 1155 Ma and, outside of local shear zones, are minimally deformed (McLelland et al., 1988a). Throughout the region, tectonism and metamorphism peaked at ca. 1180–1160 Ma. A major feature of the Shawinigan accretion is the large Maberly shear zone (MSZ, Fig. 5), along which the Frontenac domain was thrust to the northwest for large but unspecified distances, making the Shawinigan a two-sided orogen. Final displacement on the Maberly thrust has been dated by U-Pb zircon TIMS analysis at 1156 ± 2 Ma (Rivers and Corrigan, 2000; Corfu and Easton, 1997). Whatever subsequent heating took place was likely due to the large anorthosite-mangerite-charnockite-granite complexes emplaced at ca. 1155 Ma. Notwithstanding the foregoing caveats, we will, for the sake of consistency, continue to define the Shawinigan orogeny as occurring between 1190 and 1140 Ma. This definition is also consistent with ca. 1150–1144 Ma deformation in the southern Appalachians (Southworth et al., this volume).

Shawinigan magmatism in the Adirondack Lowlands documents a pre-accretion magmatic arc that is well developed in the Adirondack Lowlands (McLelland et al., 1996; Wasteneys et al., 1999). Recently rocks, of similar age have been dated (Wodicka et al., 2003) in the vicinity of the Havre St. Pierre anorthosite (Fig. 3). At ca. 1214–1200 Ma, large quantities of calc-alkaline granitoids of the Rossie-Antwerp suite (Figs. 4 and 7C) were emplaced into the Adirondack Lowlands (Wasteneys et al., 1999; Heumann et al., 2006), heralding the onset of the accretionary ca. 1190–1140 Ma Shawinigan orogeny (Corrigan and van Breemen, 1996; Rivers, 1997; Heumann et al., 2006). The Rossie-Antwerp granitoids likely manifest upper-plate magmatism above a west-dipping subduction zone, a conclusion supported by oxygen isotope investigations (Peck et al., 2004). Shortly after ca. 1190 Ma, strong deformation and metamorphism ensued in the Adirondack Lowlands (Fig. 7D) and were accompanied by emplacement of the 1185 ± 10 Ma synkinematic, calc-alkaline Hermon Granite (Heumann et al., 2006) and 1172 ± 5 Ma Hyde School Gneiss (Wasteneys et al., 1999; McLelland et al., 1992), consisting of subequal amounts of metamorphosed pink leucogranite and tonalite (Fig. 4). Along the north shore of the St. Lawrence River, pink, leucocratic, 1172 ± 5 Ma Rockport granite is interpreted to be equivalent to the granitic facies of Hyde School Gneiss but was intruded at higher crustal levels (McLelland et al., 1992). None of these plutonic rocks has been recognized in the Adirondack Highlands, suggesting separation from the Adirondack Lowlands at that time. During this interval, the Adirondack–Mount Holly belt, which was located on the east side of the closing small ocean, was being transported toward the west.

Within the Adirondacks, SHRIMP-based geochronology demonstrates that Shawinigan metamorphism caused extensive anatexis in metapelitic rocks now represented by mylonitic migmatite (Heumann et al., 2006; Bickford et al., 2008). Dating of granitic leucosome in the migmatite yields ages of 1180–1160 Ma across the entire Adirondacks. Within the Adirondack Lowlands and southwestern Adirondack Highlands, magmatic Shawinigan zircons are devoid of later overgrowths, whereas, in the eastern Adirondack Highlands, 1050–1020 Ma overgrowths are common around 1180–1160 Ma cores. These observations are interpreted as evidence that the Adirondack Lowlands did not experience Ottawan high-temperature metamorphism (Heumann et al., 2006). This conclusion is wholly consistent with the observation that titanite and monazite from the Adirondack Lowlands do not yield ages younger than 1100 Ma and that older ages are common (Mezger et al., 1991, 1992; Heumann et al., 2006). In contrast, titanite in the Adirondack Highlands yields ages of 1030 Ma or less (Mezger et al., 1991, 1992; Heumann et al., 2006), and monazite from the region typically displays rims dated at 1050–1020 Ma. These observations suggest that post-Shawinigan temperatures in the Adirondack Lowlands remained below ~650 °C, whereas the Highlands experienced considerably higher Ottawan temperatures of ~800 °C (Kitchen and Valley, 1995; Spear and Markussen, 1997; Darling et al., 2004; Storm and Spear, 2005). Likewise $^{40}Ar/^{39}Ar$ hornblende cooling ages indicate that Ottawan temperatures did not exceed 500 °C in the Adirondack Lowlands (Streepey et al., 2000, 2001; Dahl et al., 2004). To account for this discrepancy, Mezger et al. (1992) proposed that the Adirondack Highlands and Lowlands were separated by an ocean basin and were not joined by accretion until after the Ottawan orogeny. However, this is inconsistent with the existence of ca. 1155 Ma anorthosite-mangerite-charnockite-granite stitching plutons across the entire region. An alternative sequence of events is presented next.

As noted already, Adirondack Lowlands zircons exhibit no Ottawan overgrowths, and, aside from local shear zones, titanite and monazite ages exceed 1100 Ma and are generally on the order of 1150–1170 Ma. In contrast, Highland titanite ages are ≤1030 Ma, and in the eastern Adirondack Highlands, zircons contain Ottawan overgrowths. These observations imply that during the Ottawan phase of the Grenvillian orogeny, the Adirondack Lowlands did not experience Ottawan temperatures exceeding ~650–700 °C. Hornblende $^{40}Ar/^{39}Ar$ dating confirms the zircon results and indicates Adirondack Lowlands temperatures ≤500 °C during the Ottawan phase of the Grenvillian orogeny (Streepey et al., 2000). The only consistent way to produce these results is to thrust the Adirondack Lowlands over the Adirondack Highlands during the Shawinigan collision (Fig. 7D), as indicated by mylonitic fabrics and kinematic indicators in the southern Carthage Colton zone (McLelland et al., 1996; Wasteneys et al., 1999; Baird et al., 2008; Baird, 2008). Perched on top of the Adirondack Highlands, the Adirondack Lowlands experienced some Ottawan-phase deformation but escaped the granulite-facies conditions at midcrustal depths, i.e., they became part of the orogenic lid (Rivers, 2008). Toward the end of the Ottawan phase (ca. 1045 Ma), and during orogens collapse, the Adirondack Lowlands were displaced downward to the west along the reactivated Carthage-Colton zone and juxtaposed against the granulite-facies Adirondack Highlands (Selleck et al., 2005), thus corresponding to Rivers (2008) extensional collapse model of the orogenic lid and providing a detailed example of processes associated with it.

Shawinigan metamorphism is now known to have affected the Adirondack Highlands at high metamorphic grade. However, this metamorphism has been strongly overprinted by Ottawan-phase granulite-facies recrystallization and was not recognized on a regional scale until zircon SHRIMP dating of anatectites in metapelites (Heumann et al., 2006; Bickford et al., 2008). A similar overprinting may have taken place in the Canadian Grenville Province and could account for the apparent absence of Shawinigan metamorphic effects outside of the allochthonous monocyclic belt. Note, however, that Wodicka et al. (2003) reported zircon ages of ca. 1180–1170 Ma for plutons in the vicinity of the Havre–St. Pierre anorthosite massif. The veiling of Shawinigan effects could also be due to Ottawan stacking of thrust sheets in the Allochthonous Polycyclic Belt. Another possibility is that the path taken by Adirondis during the closure of the Central metasedimentary belt was subparallel to the southeastern margin of Laurentia (Fig. 7D), thus precluding the effects of a direct collision with central and eastern Laurentia and minimizing deformation and metamorphism. In contrast, the Shawinigan collisions in the southern Appalachians, as in the Adirondacks (Fig. 7D), experienced a more head-on encounter (Tohver et al., 2006), the effects of which were more intense and widespread.

Ca. 1155 Ma Anorthosite-Mangerite-Charnockite-Granite (AMCG) Suite

Following Shawinigan contraction, the eastern half of the Grenville Province was intruded by voluminous magmas of the ca. 1155–1140 Ma anorthosite-mangerite-charnockite-granite (AMCG) suite (Figs. 1, 3, and 7E). Emslie (1978) summarized the field relations of massif anorthosite and drew attention to the observation that, whereas these bodies commonly occur in high-grade terranes, they rarely show evidence of being synorogenic. Accordingly, anorthosites and their associated mangerite-charnockite granitoids became regarded as “anorogenic” (Emslie, 1978, 1985; Anderson, 1983; Hoffman, 1989; McLelland, 1991). Although the term “anorogenic” was never clearly defined, it implied a setting of mild extension in which rising gabbroic magmas differentiated at the crust-mantle interface and evolved slowly under anhydrous, quiescent conditions and pressures of 10–12 kbar (Emslie, 1978, 1985). It was proposed that rifting was sufficiently slow to ensure that the ponded magmas remained at mid- to deep-crustal sites for extended periods and that absence of contraction precluded large-scale mixing with crustal materials to produce hybrid magmas. Such hybridization has been documented in this sort of setting by Rivers (2008) and Grammatikopoulos et al. (2007), who interpreted ca. 1165–1175 Ma (U-Pb zircon TIMS) plutons of the Frontenac terrane as the result of mixing between mantle melts and crustal materials to produce a monzonite-diorite-gabbro-syenite suite. According to Emslie (1978, 1985), an "anorogenic" environment involved thick, tectonically stable crust that could provide the hot, high-pressure, tectonically stable, and anhydrous conditions required by petrological models keyed to deep, protracted fractionation of gabbroic magmas. Associated mangerite-charnockite granitoids were interpreted as anatectic melts produced by fusion of deep continental crust (Emslie, 1978) and emplaced with the anorthosites into the middle or upper crust (Valley et al., 1990) as anorthosite-mangerite-charnockite-granite suites. Some investigators (e.g., Hoffman, 1989) suggested that subcontinental hotspots and plumes meet the conditions necessary for "anorogenic" magmatism, but this model has not been conclusively demonstrated in the field.

A rapidly expanding database of zircon geochronology within the Grenville Province (e.g., Rivers, 1997) has shed light on anorthosite-mangerite-charnockite-granite petrogenesis. By 1996, it had become evident that many anorthosites were intruded during intervals similar in age to nearby contractional orogenesis (McLelland et al., 1996; Corrigan and Hanmer, 1997; Morrisett et al., 2009; Hamilton et al., 2004). McLelland et al. (2004) utilized U-Pb SHRIMP zircon dating to tightly constrain Adirondack anorthosite-mangerite-charnockite-granite ages to average values of 1154 ± 6 Ma for Marcy anorthosite (n = 13) and 1158 ± 5 Ma for associated mangerite-charnockite granitoids (n = 13). Not only do these ages demonstrate the coeval nature of the bimodal suite, but, even more importantly, they establish emplacement of the Adirondack anorthosite-mangerite-charnockite-granite suite during terminal Shawinigan orogenesis. This timing coincides with geon 11 emplacement ages of other large anorthosite-mangerite-charnockite-granite massifs in southeast Laurentia (e.g., Lac St. Jean), suggesting that effects of the Shawinigan accretion were widely felt. Possible mechanisms for anorthosite-mangerite-charnockite-granite suite production in this sort of setting include slab rollback (Mosher et al., 2008), mild extension in a backarc basin (Rivers, 2008), lithospheric-scale shear zones (Myers et al., 2008), and lithospheric delamination (McLelland et al., 2004; Morrisett et al., 2009). Each of these can bring asthenosphere into proximity with the crust, resulting in ponded gabbroic magmas, elevated temperatures, and quiescent differentiation. However, our preference is for the delamination-related model, because it is the most consistent with the late- to postorogenic setting indicated by geochronology. Accordingly, our discussion is built around this model, although the petrogenetic arguments could be equally well applied to appropriate examples based upon the alternative models cited above.

Shawinigan granulite-facies assemblages in the Adirondack Highlands (McLelland et al., 1988b) suggest overthickening of the crust during the ca. 1190–1170 Ma accretion of the Highlands–Mount Holly belt to Laurentia, and it follows that the underlying dense, cold lithosphere also thickened downward into asthenosphere. Subsequently, this dense lithospheric keel is inferred to have undergone delamination either by foundering or by convective erosion of the thermal boundary layer (Turner et al., 1992; England and Platt, 1994). This was followed by ascent of asthenosphere toward the base of the crust and ensuing rapid uplift resulting in dynamic equilibrium between buoyancy and contractional forces. Shawinigan contraction appears to have terminated by ca. 1160 Ma based upon truncation of ca. 1190–1172 Ma fabrics in Frontenac plutons by the ca. 1160 Ma Kingston dikes (Davidson, 1998) and by the minimally deformed state of scattered anorthosite-mangerite-charnockite-granite plutons in the Adirondack Lowlands.

The dynamically balanced orogen provides "anorogenic" conditions of depth, heat, low H_2O activity, and tectonic quiescence considered to be essential for anorthosite genesis, but it does so within the context of Shawinigan orogenesis rather than an "anorogenic" setting. Following delamination, asthenosphere ascended to fill the potential void left by removal of the dense keel and underwent depressurization melting to produce tholeiitic basalt (probably of high-alumina affinity) that ponded at the density lid of the crust-mantle interface. At pressures of 10–12 kbar, the gabbros fractionated olivine + pyroxene, both of which sank back into the mantle, whereas plagioclase cumulates floated and formed crystal-rich mushes (Fram and Longhi, 1992). Repeated pulses of this process produced large volumes of coarse plagioclase, and associated fractionated melts were likely to have been of leucogabbroic composition. Simultaneously, heat of crystallization, as well as advected heat, caused widespread partial melting of lower crust to produce dry, quartz-poor magmas (Wyllie, 1975), some of which commingled with gabbroic magmas to produce elevated initial strontium isotope ratios and high potassium concentrations in the anorthosites (Dymek and Seifert, 2008). In addition, lower crustal pyroxene + plagioclase restites reacted with gabbroic melts to further enhance mild crustal trace-element signatures, as well as extend the life of plagioclase on the liquidus (Emslie et al., 1994). Eventually, crustal weakening and compressional relaxation led to ascent of the mangerite-charnockite granitoids and leucogabbroic magmas. The latter disrupted and transported meter-scale rafts of coarse anorthosite cumulates containing high-pressure pyroxene. Following emplacement, interstitial magmas continued to crystallize adcumulate plagioclase, leaving a more pure anorthosite and an increasingly silica-poor, iron-titanium–rich residual ferrodioritic magma that eventually was filter-pressed into fractures in the anorthosite. The resulting anorthosite-mangerite-charnockite-granite suite is found in the Adirondack Lowlands and the Frontenac terrane, indicating that these were connected to the Adirondack Highlands by ca. 1155 Ma. Bickford et al. (2008) utilized Hf isotopic compositions from zircons in Marcy anorthosite and associated gabbro to show that the parental magmas of the anorthosites were produced from enriched mantle.

Other anorthosite-mangerite-charnockite-granite occurrences display similar late- to post-tectonic emplacement relative to regional geology. Within the Grenville Province, these include geon 11 anorthosite-mangerite-charnockite-granite suites (Figs. 1 and 3) such as the Morin (1155 ± 3 Ma; Doig 1991; Martignole, 1996), the Lac St. Jean (1170–1145 Ma; Hebert and van Breemen, 2004; Higgins and van Breemen, 1992, 1996), the Atikonak River (1133 ± 10 Ma; Emslie and Hunt, 1990), and parts of the Sheldrake lobe of the Havre–St. Pierre (dated at ca. 1129–1150 Ma by U-Pb zircon geochronology; Wodicka et al., 2003). We propose that the first two have the same relationship to the Shawinigan orogeny as the Adirondack anorthosite-mangerite-charnockite-granite suite. Field evidence indicates that at least part of the Lac St. Jean anorthosite was emplaced simultaneously with strike-slip displacement on a large, vertical shear zone at 1157 ± 3 Ma (Higgins and van Breemen, 1992, 1996). The ca. 1145 Ma ages for the Lac St. Jean, as well as ages for Atikonak and much of the Havre–St. Pierre anorthosites, were obtained on associated granitoids (Emslie and Hunt, 1990), and further direct dating of anorthosite is needed.

In Norway, the ca. 1100–930 Ma Sveconorwegian orogen involved subduction, collision, high-grade metamorphism, and plutonism. The large Rogaland anorthosite-mangerite-charnockite-granite complex was emplaced rapidly at 930 ± 3 Ma (Bingen et al., 1993, 2005; Scharer et al., 1996; Duchesne et al., 1999), an age that coincides with the end of regional contraction in the area. These workers agree that the anorthosite-mangerite-charnockite-granite plutonism occurred at the end of, and outlasted, tectonism associated with the Sveconorwegian orogeny. This relationship suggests that the magmatism took place during the dynamically balanced phase of the waning Sveconorwegian orogeny, similar to the model described for the Adirondacks.

Another example of the late- to post-tectonic setting of anorthosite-mangerite-charnockite-granite magmatism is provided by ca. 1060–1050 Ma anorthosites and granitoid suites (e.g., Parc des Laurentide area [Fig. 5]; Lac Allard, Riviere Magpie, and Romaine Riviere lobes of the Havre–St. Pierre massif [Fig. 3]) that directly postdate peak Ottawan orogenesis in the Grenville Province (Higgins and van Breemen, 1992; Corrigan and van Breemen, 1996; Morrisett et al., 2009). In addition, there is the relatively undeformed ca. 1050–1020 Ma CRUML (Chateau-Richer, St. Urbain, Mattawa, and Labrieville) belt of alkalic anorthosites (Dymek and Owens, 1998; Owens and Dymek, 2001; Owens et al., 2004) between L (Labrieville) and SU (St. Urbain) in Figure 3. There are other members of the group, including the southwest lobe of the Havre–St. Pierre anorthosite (1062 ± 4 Ma; Morrisett et al., 2009). Compositionally similar alkalic anorthosites also occur in the southern Appalachians and include the late- to postorogenic Montpelier and Roseland massifs, which have been dated at ca. 1050 Ma (Aleinikoff et al., 1996, 2004; Owens et al., 1994; Owens and Samson, 2004).

Hawkeye Granite Event (1106–1093 Ma)

Following emplacement of the anorthosite-mangerite-charnockite-granite suite, little is recorded in the Grenville Province for almost 60 m.y., except for local contractional events at ca. 1120 Ma (Wodicka et al., 1996). The only other activity appears to have been the filling of the Flinton basin in the Mazinaw terrane (Fig. 7E) with fluvial clastics and shallow-marine carbonates that were deposited on top of eroded Elzevirian plutons and sediments (Moore and Thompson, 1980). The presence of ca. 1155 Ma detrital zircons (Sager-Kinsman and Parrish, 1993) fixes a maximum age for the sedimentation. The minimum age is fixed by ca. 1080 Ma Ottawan metamorphism in the Mazinaw terrane (Corfu and Easton, 1995) that affected the Flinton sediments. The absence of Archean detrital zircon grains suggests that the provenance of the basin was restricted. Within the Adirondacks, scattered exposures of deformed hornblende granite with U-Pb zircon TIMS crystallization ages of 1106–1093 Ma (TIMS) define the Hawkeye granite suite and indicate a magmatic episode that is not recognized elsewhere in the Grenville Province. The tectonic significance of this suite remains unclear, although its age coincides with major mafic magmatism in the Midcontinent Rift (Davis and Green, 1997).

The culminating events of the tectonic evolution of the Grenville Province were deformation and metamorphism due to the Ottawan phase of the Grenvillian orogeny (1090 –1020 Ma; Rivers, 2008), including orogen collapse at ca. 1020 Ma (Fig. 7F). These events were followed by the Rigolet phase (1010–980 Ma), which involved renewed contraction and further collapse (Rivers, 2008).

Grenvillian Orogeny: Ottawan (1090–1020 Ma) and Rigolet Phases (1010–980 Ma)

Canadian Grenvillian Orogen

During the Ottawan orogenic phase (1090–1020 Ma), most of the polycyclic allochthons of the Grenville Province were thrust northwest toward, and over, the Archean and Paleoproterozoic foreland (Fig. 7F). This contraction was renewed during the less intense 1010–980 Ma Rigolet orogenic phase. Both orogenic phases included extensional faulting that was both synchronous with and postdated thrusting. The interplay between these two modes of displacement facilitated exhumation of deep, high-grade terranes by (1) thrusting at their leading front and normal faulting at their rear, and (2) late extensional collapse of large tracts of the orogen. Representative cross sections of the final architecture for the western, central, and eastern Grenville Province are shown in Figures 8A–8D. Note that in these cross sections, some faults are marked with both reverse and normal arrows to reflect the important roles played by each. Tectonic style was also determined by the presence of crustal ramps composed of cold, rigid crust of the Archean craton that served as avenues for transport of hot, ductile thrust sheets to shallow depth where they moved to the northwest along footwall flats (Figs. 8B and 8D). Very high pressures (14–18 MPa, ~800 °C) are recorded in thrust sheets from the base of the hanging wall of the Ottawan allochthons (Indares et al., 2000; Krauss and Rivers, 2004). There also exist allochthonous and parautochthonous high-pressure belts containing eclogite and eclogitic remnants recording crustal conditions of 1700–1900 MPa and 800–900 °C, corresponding to depths of 50–60 km at the base of nearly doubly thickened orogenic crust. Three areas of allochthonous high-pressure rocks are shown in black with white dots in Figure 3, and a fourth forms the Algonquin–Lac Dumoine domain in Figure 6 (Bussy et al.,

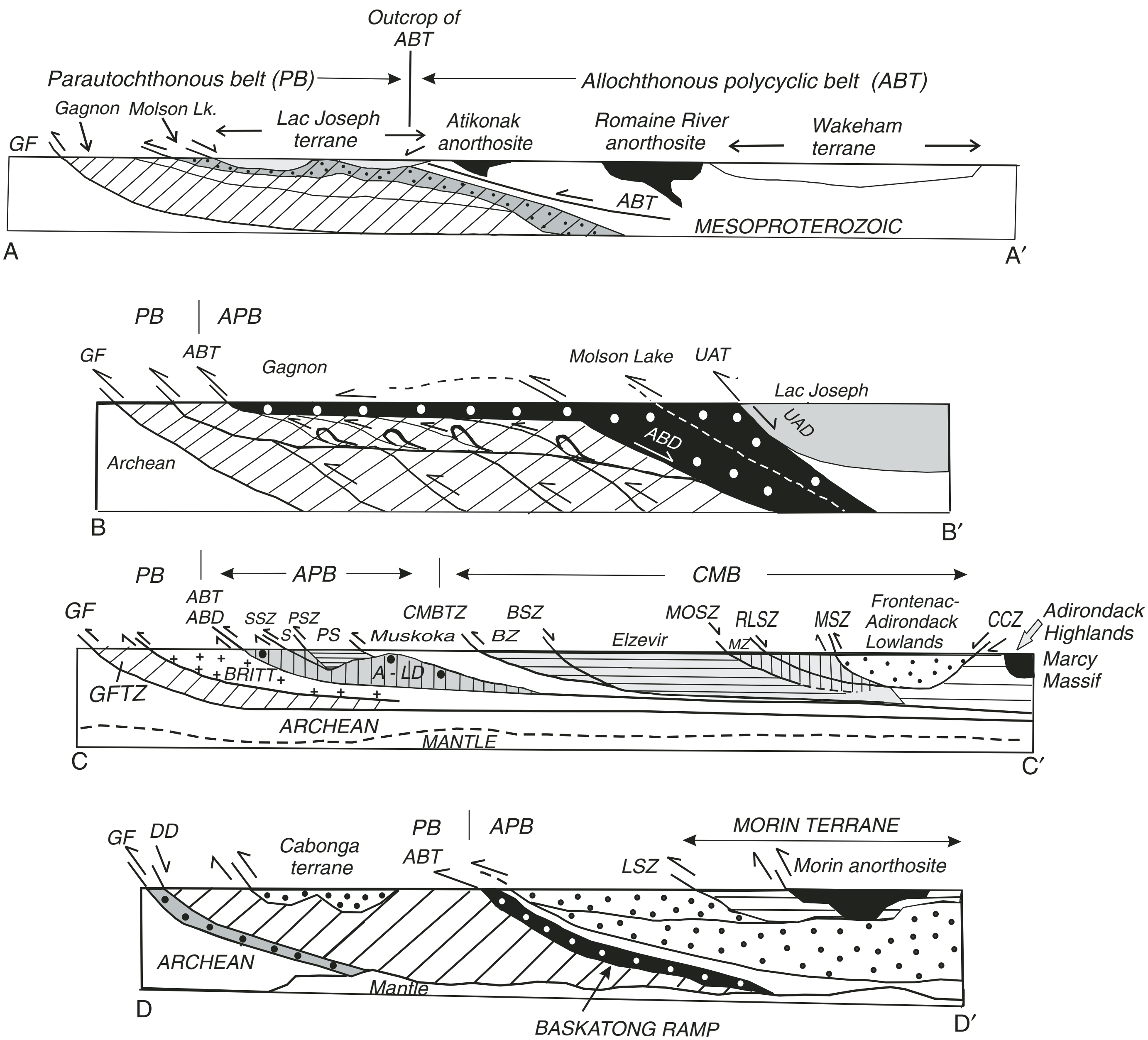

Figure 8. Schematic cross sections (not to scale) through the Grenville Province. Patterns are the same as those used on Figure 2.The Allochthonous Polycyclic Belt (APB) lies south of the Allochthon Boundary thrust (ABT), and the Parautochthonous Belt (PB) lies north of it. (A–A′) Cross section from the Grenville Front (GF) to the Wakeham terrane. The high-pressure assemblages (gray with black dots) sit below the Allochthon Boundary thrust and are of Rigolet age (modified from Rivers, 2008). (B–B′) Cross section including the Gagnon terrane Ottawan high-pressure zone where the allochthonous high-pressure (aHP) belt (black with white dots) is thrust over the Parautochthonous Belt along the Allochthon Boundary thrust, transferring Labradorian granitoids up a ramp and onto the foreland flat. The overburden load developed at the base of the Allochthon Boundary thrust hanging wall resulted in eclogite. Normal detachment faults (ABD—Allochthon Boundary detachment; UAD—upper allochthon detachment) enhanced exhumation of high-pressure assemblages (modified from Davidson; 1998, Hynes et al., 2000; Rivers, 2008). (C–C′) Cross section of the Grenville Province from Georgian Bay to the Adirondack Highlands. The high-pressure belt with retrograde Ottawan eclogite (black dots) lies well above the Allochthon Boundary thrust (modified from Carr et al., 2000; Rivers, 2008). (D–D′) Seismically constrained cross section in western Quebec. The Ottawan aHP (black with white dots) lies above the Allochthon Boundary thrust, whereas the Rigolet parautochthonous high-pressure belt lies below it (modified from Martignole et al., 2000; Rivers, 2008). Abbreviations: A-LD—Algonquin–Lac Dumoine domain; BR—Britt domain; BT—Bancroft terrane; BSZ—Bancroft shear zone; CCZ—Carthage-Colton shear zone; CMB—Central metasedimentary belt; CMBTZ—Central metasedimentary belt thrust zone; DD—Dorval detachment; GF—Grenville Front; GFTZ—Grenville Front tectonic zone; LDT-LSZ—Labele shear zone; MZ—Mazinaw domain; MOSZ—Morton shear zone; MSZ—Maberly shear zone; M—Muskoka domain; MT—Morin terrane; PS—Parry Sound domain; PSZ—Parry Sound shear zone; SSZ—Shawanaga shear zone at base of Shawanaga domain.

1995; Rivers et al., 2002; Ketchum et al., 1998). In the northeastern segments of the Grenville Province, eclogite assemblages have been dated at ca. 1090 Ma (Indares et al., 2000; Ketchum and Krogh, 1998). In both regions, 1085–1050 Ma granulite-facies assemblages overprint the eclogite (Rivers, 2008). Within the Parautochthonous Belt, high-pressure belts were formed at ca. 1005–980 Ma, i.e., a timetable that correlates the Rigolet orogenic phase.

Rivers (2008) compiled geochronological and pressure-temperature data from across the Grenville Province, and proposed a collisional model of the Grenvillian orogen as a large (~1500 × 800 km), long-term (1090–980 Ma), thick (50–60 km), hot (800–900 °C at middle crust) orogen characterized by a wide (>600 km) plateau in its hinterland and cold, rigid Archean and Paleoproterozoic crust in its foreland, suggesting that it was similar in type and scale to the present Himalayan orogen. A principal factor in Rivers' model is the high ductility of the hot middle crust. Jamieson et al. (2007) developed numerical models that produce the same hot, ductile middle crust. This condition results in extensive partial melting and channel flow, together with displacement of large, hot nappes toward the foreland, such as the far-traveled Parry Sound nappe (Jamieson et al., 2007). Conditions existing in the Grenvillian orogen led to extreme ductility, the production of extensive migmatites, subhorizontal nappes, and sheath folds together with intense L >> S lineations parallel to the direction of vergence as well as the axes of rotated early fold axes.

On the basis of these data, Rivers (2008) provided a synthesis of the architecture and evolution of the Grenvillian orogen from 1090 to 980 Ma. Pressure and temperature data assembled from across the orogen are used to subdivide it into high- (H = 1400–1900 MPa), medium- (M = 1000 ± 200 MPa), and low- (L = 400–600 MPa) pressure belts. These belts are shown in Figure 3, where the prefix a or p is used to signify Allochthonous (a) or Parautochthonous (p) Belts. The data indicate that peak Ottawan metamorphic temperatures of the allochthonous, midcrustal segments of the allochthonous medium- to low-pressure belt were relatively uniform across the orogen at 800–900 °C, with most pressures in the range 800–1100 MPa, corresponding to depths of 25–30 km in a doubly thickened crust. There are also a few low-pressure belts with pressures <600 MPa (<18 km), and more may be delineated as mapping continues, hence the inclusive designation allochthonous medium- to low-pressure. Within this broad belt, local differences between timing of events in separate segments exist due to local variations over an ~40 m.y. interval. Most allochthonous medium-pressure segments formed at 1090–1050 Ma, whereas the allochthonous low-pressure segments were formed at 1050–1020 Ma. The former originated during Ottawan thrusting, and the latter originated when late Ottawan extension dropped the low-pressure rocks down into contact with warmer blocks of midcrust. The extension was largely accommodated on the former Allochthon Boundary thrust surface and is referred to as the Allochthonous Detachment fault (ADF, Fig. 3). Calculated cooling rates for the hot allochthonous middle-pressure rocks exceed those for the already cooler low-pressure rocks.

Current levels of erosion in the Allochthonous Polycyclic Belt generally expose rocks that crystallized at 20–25 km depth and that represent the hot middle crust of the Grenvillian orogen (allochthonous medium- to low-pressure) However, near each end of the belt, there exist large tracts of crust that did not experience Grenvillian penetrative deformation or high-grade metamorphism (Rivers, 2008). In the southwest, these rocks occur within the accreted terranes of the Central metasedimentary belt (Figs. 3 and 5), where they contain Elzevirian (1245–1225 Ma) and/or Shawinigan (1190–1140 Ma) assemblages of low to medium metamorphic grade. In the northeast, they are found within the large region shown in gray (Fig. 3) that consists of mainly Labradorian (ca. 1670–1620 Ma) high-grade gneiss that lacks any significant Grenvillian overprint. Following Laubscher's (1963) terminology in the Alps, Rivers (2008) designated these regions as remnants of an orogenic lid that once formed the suprastructure above the hot, ductile middle crust of the Grenvillian hinterland. According to the model, after 40 m.y. (1090–1050 Ma) of contraction and heating, the tectonic plateau became increasingly unable to support its own weight, and the effects of both gravitational instability and waning contraction promoted an extensional collapse that peaked at ca. 1020 Ma. This led to the formation of a crustal-scale, orogenwide horst and graben terrane, within which the orogenic lid was downfaulted while the hot midcrustal portions were exhumed in horsts as core complexes and gneiss domes. Where exposed, the boundaries between the orogenic lid remnants and the allochthonous medium- to high-pressure terranes consist of normal faults consistent with the ca. 1020 Ma extensional collapse model (Rivers, 2008).

The orogenic lid records a complex orogenic history in the Central metasedimentary belt, where it includes the accreted terranes within the heavy dark outline in Figure 3 (Rivers, 2008). The following domains lie within this area: Mont Laurier, Cabonga, Frontenac, Sharbot Lake, and Adirondack Lowlands. These are characterized by the same absence of Ottawan penetrative metamorphism as found in the northeastern orogenic lid, and similar slow cooling patterns exist. To the northwest, the accreted terrane(s) of the Central metasedimentary belt thrust zone (CMBTZ, Fig. 5) shows an Ottawan parautochthonous medium- to low-pressure belt overprint on rocks that are a mixture of several varieties of crust stacked together by Elzevirian and Ottawan thrusting. In the southeastern part of the zone, the Mazinaw domain (Fig. 5) records high-grade metamorphism at ca. 1050–1000 Ma (Corfu and Easton, 1995) and may represent a horst of Ottawan allochthonous low pressure (Rivers, 2008). Strong Ottawan overprints are also present in the accreted, allochthonous high- to low-pressure monocyclic rocks of the Adirondack Highlands and Morin terranes.

The parautochthonous belt contains parautochthonous medium- to low-pressure amphibolite- and granulite-facies assemblages similar to those in the allochthonous medium,- to high-pressure belt, but there are major differences in ages of metamorphism that occurred at 1005–980 Ma in the parautochthon, i.e., 15–40 m.y. years younger than in the allochthonous belts.

Rivers (2008) defined two segments with parautochthonous high-pressure Rigolet characteristics that occur in the parautochthon. The largest of these is in the Gagnon terrane and is in contact with the northeastern orogenic lid (Figs. 3 and 8). Here, at 1005–980 Ma, the parautochthonous fold-and-thrust belt experienced Grenvillian pressures ranging from 600 to 1400 MPa (16–30 km) and temperatures from 450 to 800 °C. Within the overlying parautochthonous high-pressure segment, pressures ranged from 1400 to 1800 MPa (40–50 km), and temperatures ranged from 750 to 850 °C. In the Grenville Front tectonic zone (Fig. 3), similar temperatures prevailed, but pressures were in the 600–1500 MPa range, with hornblende cooling ages of 905–975 Ma. In these areas, the allochthonous high-pressure and parautochthonous high-pressure segments are in contact across the allochthon boundary thrust, and at 1090–1050 Ma, the allochthonous high-pressure assemblages were thrust for over 100 km into their present position. In contrast, the parautochthonous high-pressure assemblages were formed when contraction resumed in the Rigolet phase and moved the cooler and more rigid Allochthonous Polycyclic Belt over its foreland at 1005–980 Ma. Reverse faulting in the footwall PB displaced parautochthonous high-pressure assemblages to the northwest, where they were subsequently exhumed by orogen collapse at ca. 980–950 Ma, marking the termination of the Grenvillian orogeny (Rivers, 2008).

Geon 10 plutonic rocks are of limited occurrence in the Canadian Grenville Province. The most extensive tract consists of undeformed 1060–1050 Ma charnockites and granites in the Parc des Laurentides area of Quebec (Fig. 5). These have been dated by U-Pb zircon single-grain analysis (Higgins and van Breemen, 1992, 1996; Corrigan and van Breemen, 1996). They are part of the Chateau-Richer, St. Urbain, Mattawa, and Labrieville belt of anorthosite-mangerite-charnockite-granite magmatism discussed previously. A monzonite pluton dated by the single-grain method yielded an age of 1126 ± 7 Ma (Emslie and Hunt, 1990) and is associated with ca. 1060 Ma anorthosite in the southern lobe of the Havre–St. Pierre pluton (Higgins and van Breemen, 1992; Morrisett et al., 2009). Small, undeformed plutons of the ultrapotassic, alkaline Kensington-Skootamata suite, ranging from syenite to diorite and including pyroxenite and xenolith-bearing minette dikes, occur in the Mont Laurier terrane of the Central metasedimentary belt orogenic lid (Figs. 3 and 5) and were described and dated at ca. 1090–1070 Ma by U-Pb zircon multigrain analysis (Corriveau, 1990). Corriveau and van Breemen (2000) interpreted the suite to have formed during early stages of collapse of the Grenville orogen, and the older ages may reflect the analytical method used. Also present in the Mont Laurier terrane, the ca. 1060 Ma Guenette granite intruded as sheets along a major fault zone and is presumably related to late orogen collapse. Farther south in the Frontenac domain, the small Rideau, Wolfe Lake, and Westport plutons consist of monzonite and, based on U-Pb zircon multigrain age TIMS analyses, were emplaced at ca. 1075 Ma (Marcantonio et al., 1990). Similar small plutons of this age and composition are scattered through the Central metasedimentary belt (Figure 3.22 *in* Davidson, 1998,). Recently, Easton and Kamo (2008) reported the discovery of additional 1060–1050 granitic stocks in the western Central metasedimentary belt. Notwithstanding these additions, the volume of ca. 1060–1050 granitic magmatic activity in the Canadian Grenville Province is relatively small.

Adirondack Grenvillian

The Adirondacks (Fig. 4) form a southern extension of the of the Grenvillian hinterland plateau (Fig. 3). The Adirondack Highlands belong to the allochthonous medium- to low-pressure belt segment, and the Adirondack Lowlands are part of the western orogenic lid (Rivers, 2008). Ottawan metamorphic assemblages at the current surface of the Adirondack Highlands reflect crustal conditions of ~800 MPa and 800 °C (Bohlen et al., 1985; Spear and Markussen, 1997; Kitchen and Valley, 1995; Darling et al., 2004; Storm and Spear, 2005, among others), corresponding to ~20–25 km depth. Because present depth to the Moho is on the order of 40 km (Brown et al., 1983; Hughes and Luetgert, 1992), these data indicate an Ottawan thickness of 60–65 km. The kilometers-wide Carthage-Colton shear zone (CCZ, Fig. 4) dips 45° to the northwest and separates the granulite-facies Adirondack Highlands (ca. 1.35–1.05 Ga) from the amphibolite-grade Adirondack Lowlands (ca. 1.21–1.14 Ga). Normal-sense detachment faulting along the zone at 1047 ± 5 Ma juxtaposed the amphibolite-facies Adirondack Lowlands against the granulite-facies Adirondack Highlands (Johnson et al., 2004; Selleck et al., 2005). Detailed studies of fabrics and kinematic indicators along the southwestern Carthage-Colton zone demonstrate an earlier, ca. 1170 Ma reverse fault displacement, in which the Adirondack Lowlands were thrust to the southeast over the Adirondack Highlands (Baird, 2008; Baird et al., 2008). Johnson et al. (2004) interpreted the northern segments of the zone as an early oblique thrust followed by ca. 1050 Ma extensional collapse. The steep Black Creek shear zone (BCSZ, Fig. 4), south of the St. Lawrence River, separates the Adirondack Lowlands from the Frontenac terrane (Heumann et al., 2006). Most fabrics in the Black Creek shear zone are consistent with late, low-grade deformation. Significantly, the undeformed ca. 1160 Ma Kingston dike swarm (Davidson, 1998), which cuts ca. 1160 Ma plutons in the Frontenac terrane, does not cross this zone, and Heumann et al. (2006) suggested that the zone may be continuous with late deformation within the Labelle shear zone, which forms the western margin of the Morin terrane (Fig. 5). In addition, the bedrock of the Frontenac terrane is dominated by quartzites and metavolcanics, whereas marbles and plutons dominate the Adirondack Lowlands, and the differences suggest some offset between the two terranes. Lineations in the Black Creek shear zone are subhorizontal and plunge to the south, suggesting oblique sinistral displacement.

The ca. 1090 Ma onset of the Ottawan phase is not tightly constrained in the Adirondacks; however, the peak of metamorphism itself is well dated by SHRIMP analysis of zircon rims, as well as equidimensional multifaceted metamorphic zircons, which yield ages that peak strongly at ca. 1050 Ma (McLelland et al., 2001, 2004; Heumann et al., 2006). Monazite dating by electron microprobe yields metamorphic age peaks at 1050–1020 Ma (Williams et al., 2008), and titanite ages are in the range ca. 1030–980 Ma (Mezger et al., 1992; Heumann et al., 2006). These results confirm a strong granulite-facies pulse of metamorphism at ca. 1050 Ma, and show that this is the age of peak Ottawan metamorphic conditions.

The Adirondack Highlands are structurally characterized by intense ductile deformation and widespread penetrative fabrics with strong, subhorizontal tectonic foliation and gently plunging ribbon lineations (L >> S) that trend ~E-W. The earliest mappable folds consist of kilometer-scale recumbent (F_1) isoclines (e.g., CLI, Fig. 4) having axes oriented parallel, or nearly parallel, to the ribbon lineations. These folds wrap earlier tectonic foliations as well as intrafolial isoclines (F_0), attesting to multiple episodes of deformation (McLelland and Isachsen, 1986). Very large ~E-W open, upright folds (F_2) refold the isoclines and have axial traces that can be mapped for over 100 km (P, Fig. 4). The axes of the F_2 folds are parallel to the ribbon lineations as well as the F_1 fold axes, reflecting the influence of a common major event late in the area's history. The latest fold set (F_3) is approximately normal to the F_2 fold axes and consists of broad, open, upright flexures of NNE orientation. Intersections between the threefold sets result in interference patterns including basin and dome structures. The F_1–F_3 fold sets are demonstrably of Ottawan origin because their effects are present in the ca. 1155 Ma anorthosite-mangerite-charnockite-granite rocks as well as the ca. 1100 Ma Hawkeye granite (Chiarenzelli and McLelland, 1990). F_2 and F_3 folds, but not the isoclinal nappe-like F_1 folds, affect the ca. 1050 Ma Lyon Mountain granite. This constrains the timing of F_2 and F_3 folding to the 1050–1040 Ma early extensional interval of the Ottawan orogeny. Multiple deformation and fold interference domes and basins are also present in the Adirondack Lowlands, with four fold sets identified by Weiner et al. (1984); however, due to the lack of Ottawan deformation in the lowlands, it is not clear how these correlate with fold sets in the highlands.

The only igneous unit associated with the Ottawan phase in the Adirondacks is the widespread and distinctive Lyon Mountain granite (LMG, Fig. 4), consisting mainly of pink hypersolvus leucogranite and subordinate charnockite (Postel, 1952; Whitney and Olmsted, 1988; McLelland et al., 2002; Valley et al., 2009). It was intruded principally as plutons and conformable sheets of variable thickness that postdate F_1 isoclinal folds but are variably affected by upright F_2 and F_3 folds. Grain shape fabric is uncommon except at contacts and within shear zones. Potassic and sodic alteration are related to regional hydrothermal fluids driven by heat from granitic plutons during late magmatism. These fluids were also responsible for hydrothermal deposition of Kiruna-type magnetite ores accompanied by the sodic alteration (Foose and McLelland, 1995; McLelland et al., 2002; Valley et al., 2009). SHRIMP and single-grain methods have been used to date a total of 13 samples of Lyon Mountain granite (Table DR1 [see footnote 1]) that yield a tight array of ages from 1058 to 1035 Ma (McLelland et al., 2001, 2002). These results place intrusion of the granite within the interval of peak Ottawan metamorphism and orogen collapse. Lyon Mountain granite is locally folded by upright, open F_2 and F_3 structures, an example of which can be seen in the west-central Adirondack Highlands, where the large, upright E-W, F_2 Arab Mountain anticline (A, Fig. 1) causes the granite's outcrop pattern to make a deep reentrant from west to east and back again to the west along its southern limb. This relationship between F_2 and F_3 folding and Lyon Mountain granite constrains the timing of this folding to near the termination of Ottawan orogenesis.

Selleck et al. (2005) demonstrated that 1047 ± 5 Ma Lyon Mountain granite intruded synkinematically during down-to-the-west detachment faulting of the Adirondack Lowlands along the Carthage-Colton zone and suggested that emplacement of the granite enhanced orogenic collapse. The outcrop pattern of Lyon Mountain granite is approximately peripheral to the Adirondack Highlands (Fig. 4), and, given the interplay between it and orogen collapse, Selleck et al. (2005) suggested that the eastern Adirondacks might also contain late Ottawan detachment faults. Field research between Willsboro and Fort Ann (Fig. 4) has revealed a wide (~10–15 km) eastern Adirondack shear zone with straight gneiss (L >> S) and mylonite dipping (15–45°) to the east. Stained slabs of 1155 ± 10 Ma anorthosite-mangerite-charnockite-granite megacrystic granite with ribbon lineation indicate that the dominant sense of shear displacement is topside down and to the east (McLelland et al., 2009). U-Pb monazite analysis by electron microprobe (Williams et al., 2008) from mylonitic samples collected near Fort Ann (Fig. 4) indicates peaks of crystallization at 1050–1030 Ma. These occur on late tips that are aligned parallel to extensional ribbon lineation formed in part by feldspar tails, and they support late Ottawan ages for the topside down-to-the-east shear sense of the kinematic indicators. Accordingly, we propose that, near the close of the Ottawan orogeny, the Adirondacks underwent orogenic collapse along both the Carthage-Colton shear zone and eastern Adirondack shear zones, resulting in a symmetrical core complex of Shuswap-scale and Shuswap-style (e.g., Teyssier and Whitney, 2002; Johnson, 2006). The Shuswap collapse took place at 58 Ma shortly after peak metamorphism at ca. 60 Ma (Teyssier and Whitney, 2002; Johnson, 2006). The short time span between peak metamorphism and collapse corresponds to the timing of similar events in the Grenvillian orogen. An additional similarity is that collapse was accompanied by intrusion of leucogranite that may have lubricated and promoted zones of weakness, thus enhancing faulting (Johnson, 2006), a feature that appears to be common in core complexes (Lister and Baldwin, 1993; Parsons and Thompson, 1993; Wells and Hoisch, 2008). Indeed, it may be that core complexes are a common feature of collapsed large, hot orogens with gravitationally unstable orogenic lids.

Corrigan and van Breemen (1996) noted that oblique normal fault displacement occurred along the Tawachiche shear zone (TSZ, Fig. 5) at ca. 1070–1050 Ma and dropped the medium-grade Montauban–La Bostonnaiss arc down into juxtaposition with the granulite-facies bedrock of the eastern margin of the ca. 1.18–1.15 Ga Morin terrane and ca. 1.37 Ga Mekinac terrane. If this shear zone is projected to the south along strike, its trace passes into the narrow Champlain Valley between the Adirondacks and the Mount Holly complex (Fig. 5), and thus it may have played a role in late extensional collapse of this region. The precise location of the master fault between the eastern Adirondack shear zone and the central Adirondack Highlands is currently unknown but is inferred to approximately coincide with the western limit of Lyon Mountain granite (dashed line in Fig. 4). We further propose, as suggested by dating of tips on oriented monazites (Williams et al., 2008; McLelland et al., 2009), that much of the ~E-W ribbon lineation of the Adirondacks developed during the ca. 1050–1040 Ma collapse of the local Ottawan orogen. In this process, large, open "a-type" elongate domes developed parallel to extension (F_2), whereas another set of "b-type" elongate domes (F_3) formed perpendicular to extension (Jolivet et al., 2004). F_1 isoclines preceded extensional collapse, and their ~E-W fold axes were drawn into near parallelism with the ~E-W extensional fabrics. The age of isocline formation is placed at 1080–1060 Ma because ca. 1050 pegmatite dikes crosscut F_1-related fabrics. The crosscutting evidence also constrains the formation of most straight gneiss (L >> S) to pre–ca. 1050 Ma Ottawan channel flow in the allochthonous medium- to low-pressure region of the Adirondacks.

Farther west, the east-dipping Bancroft shear zone (Fig. 5) was active from ca. 1020 to 935 Ma (van der Pluijm and Carlson, 1989) and records the collapse of the Ottawan orogen to the east. Also recording Ottawan collapse is the normal displacement at ca. 1050 on the east-dipping, reactivated (Davidson, 1998) Maberly shear zone. This oppositely directed collapse of the orogen was among the many fault sets that dropped the western orogenic lid into place at ca. 1040–1020 Ma.

MESOPROTEROZOIC GEOLOGY OF THE APPALACHIAN INLIERS

Recent studies of geon 11–13 terranes throughout the Appalachians, including geochronological investigations, make it possible to evaluate possible relationships between these areas and the Grenville Province. The presentation here is organized chronologically beginning with events from 1.4 to 1.22 Ga. Within each section, the organization is from north to south. Unless noted otherwise, all ages reported were acquired by U-Pb SHRIMP analysis of zircons and refer to igneous crystallization. Ages are summarized in boxes on Figure 9 and given in Ga.

Pre-Elzevirian (1.4–1.3 Ga) and Elzevirian (1245–1225 Ma) Magmatism and Metamorphism

As noted previously, the lithologic sequence of the eastern Adirondacks reappears in the Mount Holly complex, Vermont. Ratcliffe et al. (1991) and Ratcliffe and Aleinikoff (2001) utilized multigrain TIMS as well as SHRIMP analyses to determine that the oldest rocks in the Mount Holly complex are 1426 ± 9 to 1331 ± 6 Ma (Nd model age = 1.46 Ga) tonalitic, trondhjemitic, and granodioritic orthogneiss with compositions similar to those of coeval Adirondack rocks also derived from rifting of the ca. 1.4–1.3 Ga Dysart arc. The 1244 ± 8 Ma (Nd T_{DM} = 1.47 Ga) College Hill granite (Ratcliffe et al., 1991; Ratcliffe and Aleinikoff, 2001) is similar in age and composition to the Canada Lake charnockite of the Adirondacks (McLelland and Chiarenzelli, 1990). A younger 1221 ± 4 Ma granite has also been reported, and both units resemble the ca. 1250 Ma granitoids in the Central metasedimentary belt and further support regional correlation of events. The geon 14 rocks are also present in the Chester and Athens Domes (Fig. 9) along the eastern side of the Green Mountains (Karabinos and Aleinikoff, 1990; Karabinos et al., 1999, 2003; Ratcliffe and Aleinikoff, 2008). Tollo et al. (this volume) dated tonalitic gneiss in the Mount Rogers area at 1327 ± 7 Ma.

Ratcliffe and Aleinikoff (2008) reported new ca. 1325–1350 Ma ages for five granitoids within the Hudson Highlands and Manhattan Prong of New York (Fig. 9), including the Fordham gneiss. These trondhjemitic, tonalitic, and granodioritic lithologies are equivalent to those in the Mount Holly complex and are interpreted to be Dysart–Mount Holly equivalents (Ratcliffe and Aleinikoff, 2008). In addition, these authors report 1240–1243 Ma quartz-monzonite and aplite in the northern Hudson Highlands similar to the ca. 1.25 Ga granitoids in the Central metasedimentary belt. Walsh et al. (2004) dated pink granitic gneiss at 1311 ± 7 Ma in nearby western Connecticut and interpreted this unit to be a member of the Dysart–Mount Holly suite (Walsh et al., 2004, p. 749).

In the Mesoproterozoic of the New Jersey Highlands (Fig. 9), dating of orthogneiss in the calc-alkaline Losee Suite (Volkert et al., this volume) yielded ages of ca. 1.35–1.25 Ga. These results extend the ca. 1.3 Ma rifting and subsequent back-arc basin southward to at least the southern terminus of the New Jersey Highlands. Farther south, Aleinikoff et al. (2004) dated two granitoids within the Baltimore Gneiss Domes (Fig. 9) at 1247 ± 9 Ma and 1240 ± 4 Ma, both of which overlap the Elzevirian orogenic interval.

Shawinigan Orogeny (1190–1140 Ma) and Anorthosite-Mangerite-Charnockite-Granite (1160–1140 Ma) Magmatism

A Shawinigan age of 1172 ± 7 Ma has been reported from the northernmost Mount Holly complex where foliated granitic sheets crosscut gneissosity in older orthogneiss (Ratcliffe and Aleinikoff, 2001, 2008). Shawinigan ages are well represented in the Berkshire massif (Fig. 9), where dating of Tyringham granitic gneiss yields an age of 1179 ± 9 Ma (Nd T_{DM} = 1.46 Ga; Karabinos et al., 2003). This unit crosscuts all other rocks and underlies ~35% of the Berkshires. Within the Hudson Highlands, Ratcliffe and Aleinikoff (2008) obtained an age of 1174 ± 8 Ma for the widespread Storm King granitic gneiss, for which Daly and McLelland (1991) reported a Nd model age of 1.46 Ga. There is also widespread evidence of Shawinigan deformation throughout

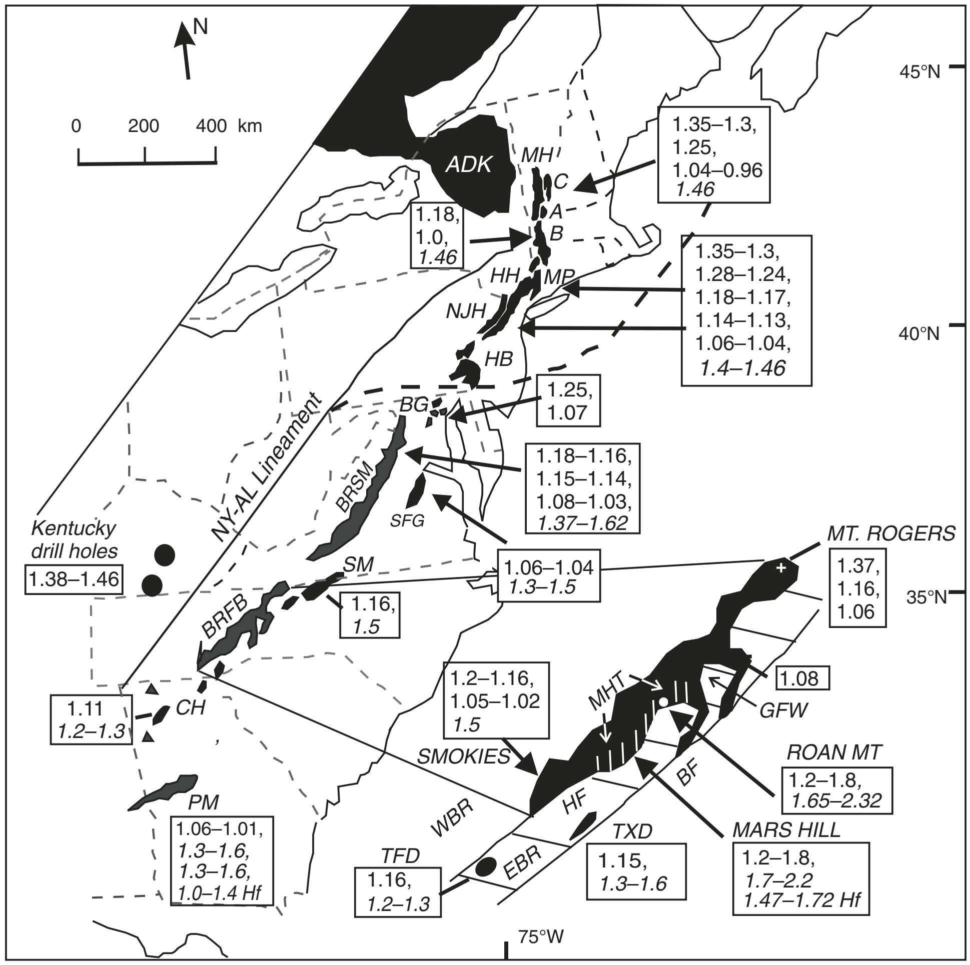

Figure 9. Generalized map of the eastern United States showing Mesoproterozoic massifs in black and their ages in boxes (T_{DM} in italics). Age references are given in text (Mesoproterozoic Appalachian Inliers section). New York (NY)–Alabama (AL) Lineament is a candidate for a Grenvillian suture zone (Zeitz and, 1978). ADK—Adirondacks; B—Berkshires; BF—Brevard fault zone; BG—Baltimore gneiss dome complex; EBR—Eastern Blue Ridge; BF—Brevard fault; BRFB—Blue Ridge French Broad massif; BRSM—Blue Ridge Shenandoah massif; CH—Corbin Hill; G—Goochland terrane; GMT—Green Mountains; GFW—Grandfather Mountain window; HF—Hayesville fault; HB—Honey Brook uplands; HH—Hudson Highlands; MHT—Mars Hill terrane (vertical white lines); MP—Manhattan Prong; NJH—New Jersey Highlands (Reading Prong); PM—Pine Mountain uplift; TFD—Tallulah Falls Dome; TXD—Toxaway Dome; WBR—western Blue Ridge. The dashed boundary represents the possible eastern limit of Laurentia at ca. 1.2 Ga (modified from Tollo et al., 2006; Carrigan et al., 2003).

the Hudson Highlands (Ratcliffe and Aleinikoff, 2008). In the New Jersey Highlands, Aleinikoff et al. (2007) reported an age of 1176 ± 11 Ma for the Byram-Hopatcong hornblende granite.

The ca. 1142 ± 8 Ma Canopus pluton and the 1134 ± 7 Ma Brewster hornblende granite gneiss of the Hudson Highlands are of anorthosite-mangerite-charnockite-granite age and have compositions characterized by ferroan, potassic granitoids (Ratcliffe and Aleinikoff, 2008; Walsh et al., 2004). These plutons crosscut older gneissic structures, mylonite zones, and isoclines, indicating a probable Shawinigan age (Ratcliffe and Aleinikoff, 2008). Currently no definitive geon 12 Ma anorthosite-mangerite-charnockite-granite–type rocks have been reported in the New Jersey Highlands; however, local sheets of anorthosite as well as charnockite suggest the presence of anorthosite-mangerite-charnockite-granite rocks (Gorring and Volkert, 2004; Volkert et al., this volume).

Within the Northern Blue Ridge (Shenandoah massif [BRSM]; Fig. 9), Southworth et al. (this volume) report new and revised ages for a large number of samples of basement units. The oldest of these is a foliated megacrystic leucogranite with an emplacement age of 1183 ± 11 Ma, and another six granitic rocks have ages ranging from 1171 ± 6 Ma to 1164 ± 8 Ma (Tollo et al., 2004) corresponding to Shawinigan events (Rivers, 1997). Three samples range from 1153 ± 9 Ma to 1143 ± 8 Ma and correlate in time with the major anorthosite-mangerite-charnockite-granite magmatism in the Adirondacks and Grenville Province. Bartholomew and Lewis (1992) reported small occurrences of anorthosite within the Blue Ridge, giving further support to anorthosite-mangerite-charnockite-granite correlation. Southworth et al. (this volume) refer to the 1183–1144 Ma collection of rocks as magmatic group I (Fig. 2). Amphibolite- to granulite-facies metamorphism and deformation of these rocks took place between ca. 1153 and ca. 1144 Ma (Southworth et al., this volume). Nd model ages range from 1.39 to 1.6 Ga and establish that none of the rocks is juvenile. In the Baltimore gneiss dome complex (Fig. 9), biotite granite intruded older gneiss at 1075 ± 10 Ma, and metamorphic overgrowths on zircons in all the older rocks are dated at ca. 1020 Ma (Aleinikoff et al., 2004).

Carrigan et al. (2003) and Ownby et al. (2004) investigated the enigmatic granulite-facies Mars Hill terrane (Fig. 9) that lies within the French Broad massif of the Western Blue Ridge and obtained ages of ca. 1.6–1.8 Ga from granulite-grade gneiss at Roan Mountain. Nd model ages for these rocks are on the order of 1.8–2.2 Ga. These Paleoproterozoic ages are so different from any others in the Appalachian inliers that they have been interpreted as signifying that much of the Mars Hill terrane represents an accreted exotic block (Carrigan et al., 2003; Ownby et al., 2004). A SHRIMP study on rocks from the Mars Hill terrane indicated the presence of many ca. 1.6–1.9 Ga zircon cores that suggest the presence of rocks of unusual antiquity in the area (Ownby et al., 2004). These authors also reported SHRIMP ages from four felsic orthogneiss units and one mafic orthogneiss from Mars Hill ranging from 1257 ± 26 Ma to 1185 ± 36 Ma, with an average of ca. 1207 Ma. Ownby et al. (2004) also concluded that by far the most significant magmatic activity in the Mars Hill terrane took place at ca. 1.2 Ga, and noted that plutons have 1.8–2.2 Ga T_{DM} ages and negative ε_{ND} at ca. 1.0 Ga. Carrigan et al. (2003) obtained a SHRIMP age of 1195 ± 19 Ma from the Cranberry Granite located just north of Mars Hill. Within the Tallulah Falls Dome of the eastern Blue Ridge (Fig. 9), Carrigan et al. (2003) and Hatcher et al. (2004) reported SHRIMP ages for the Wiley quartzofeldspathic augen gneiss at 1158 ± 19 Ma (magmatic group 1) and 1129 ± 23 Ma (magmatic group 2) for the Wolf Creek granitoid. The predominant orthogneiss of the Toxaway Dome in the eastern Blue Ridge (Fig. 9) yields SHRIMP ages of 1151 ± 17 Ma and 1149 ± 32 Ma (Carrigan et al., 2003). These ages correspond to magmatic group I of Southworth et al. (this volume) as well as anorthosite-mangerite-charnockite-granite magmatism in the Adirondacks.

The Honey Brook anorthosite in eastern Pennsylvania (HB, Fig. 9) has not been dated by zircon; however, Pyle (2004) reported that monazite dating in adjacent units constrains emplacement to >1070 Ma. The anorthosite, which is similar in composition to the Marcy massif with plagioclase in the range An50-An60 and is associated with charnockitic gneiss, underwent granulite-facies metamorphism at ca. 1050 Ma (Tarbert, 2007; Crawford and Hoersch, 1984). Sinha et al. (1996) plotted five whole-rock $^{206}Pb/^{204}Pb$ versus $^{207}Pb/^{204}Pb$ analyses that exhibit a linear correlation equivalent to a $^{207}Pb/^{206}Pb$ age of ca. 1132 Ma. This age is broadly correlative to the time of Marcy emplacement and is consistent with the fact that on a $^{207}Pb/^{204}Pb$ versus $^{206}Pb/^{204}Pb$ diagram (Fig. 10), the Honey Brook data plot below the Stacey-Kramers curve and close to Adirondack data (Sinha and McLelland, 1999). We consider the anorthosite to be of geon 11 age and possibly ca. 1155 Ma in age, but SHRIMP analysis is required.

In the Smoky Mountains of the western Blue Ridge province (Fig. 9), Aleinikoff et al. (2007) dated granitic, granodioritic, and migmatitic orthogneiss. Granitic leucosomes in the migmatite give ages of ca. 1194 Ma, and gneissic granitoids are dated at 1165 ± 7 Ma, corresponding to Shawinigan magmatism within magmatic group I. Tollo et al. (this volume) dated granitoids of the nearby Mount Rogers area at ca. 1174–1161 Ma, with associated Nd T_{DM} of 1.5–1.6 Ga. These plutons were intruded into hot (750 °C) crust at upper-amphibolite-facies, conditions reflecting regional Shawinigan metamorphism, and they are similar in age to rocks in the Shenandoah massif. The 1174–1161 Ma plutons crosscut 1327 ± 7 Ma tonalitic-granodioritic granofels at Mount Rogers.

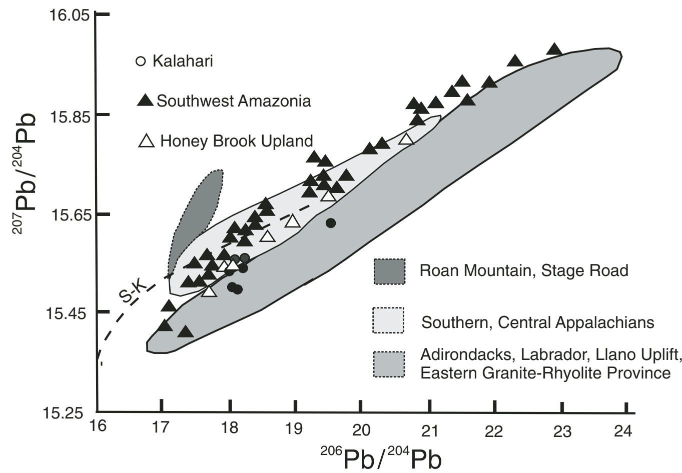

Figure 10. Plot of Pb whole-rock isotope ratios showing the Amazonian affinity of the southern and central Appalachians and the distinctly different and strongly coherent data for the Adirondacks, Llano, Labrador, and midcontinent. (Data are from Sinha et al., 1996, and Sinha and McLelland, 1999.)

Ca. 1120–1100 Ma Magmatism

A small number of granitic intrusions within the southern Appalachians yields ages similar to that of the ca. 1106–1093 Ma (multigrain TIMS) Hawkeye Granite suite of the Adirondacks (Fig. 4). These include examples from the northern Blue Ridge as well as the Great Smoky Mountain region (Fig. 3) and define magmatic group II (ca. 1127–1116 Ma) as determined by SHRIMP analysis (Southworth et al., this volume). The upper end of this range is significantly older than the Hawkeye Granite, the multigrain TIMS age of which is probably too young due to metamorphic overgrowths and thus requires redating by SHRIMP analysis. Heatherington et al. (2006) reported a secondary ion mass spectrograph (SIMS) age of 1106 ± 13 Ma (Nd T_{DM} = 1.6 Ga) for granitic gneiss from the Corbin Hill massif (Fig. 9) of the southern Blue Ridge in Georgia. In the Tallulah Falls Dome, the Wolf Creek granitic gneiss has a SHRIMP age of 1129 ± 23 Ma (Nd T_{DM} = 1.3–1.2 Ga) that overlaps both magmatic group II (Fig. 2) and anorthosite-mangerite-charnockite-granite magmatism (Hatcher et al., 2004).

Ottawan (1090–1020 Ma) and Rigolet (1010–980) Magmatism and Metamorphism

To date, the only examples of ca. 1050 Ma granitic magmatism reported from the Mount Holly complex consist of undeformed 1036 ± 6 Ma muscovite-albite pegmatites and associated sodic alteration and magnetite mineralization (Ratcliffe and Aleinikoff, 2001). These occurrences display ages, sodic alteration, and magnetite mineralizations characteristic of late Lyon Mountain magmatism in the neighboring Adirondacks. Ratcliffe and Aleinikoff (2001) reported ages of ca. 1.0 Ga for the Cardinal Brook intrusive suite in the Mount Holly complex and the adjacent Athens and Chester Domes (Fig. 9). Karabinos and Aleinikoff (1990) dated the Bull Hill Gneiss of the Mount Holly Complex at ca. 965–945 Ma, and Karabinos et al. (2003) reported ages of 1004 ± 19 Ma to 997 ± 10 Ma for alaskite dikes in the Berkshire Mountains. These authors have also reported sheets of 962 ± 1 Ma rapakivi granite.

Farther south, in western Connecticut, Walsh et al. (2004) dated deformed Danbury megacrystic granite as well as leucosomes from several migmatites. The migmatites yield ages of 1057 ± 10–1048 ± 11 Ma and date Ottawan-age anatexis. The 1045 ± 8 Ma Danbury megacrystic granite is a likely equivalent of Lyon Mountain granite (Walsh et al., 2004). Farther west in the Hudson Highlands, the Canada Hill granite pluton, with associated migmatites and pegmatites, has been dated at ca. 1014 Ma by multigrain zircon TIMS methods.

About 30 small (0.2–5 km^2) bodies of undeformed pink to gray Mount Eve granite (Gorring et al., 2004) occur within the Mesoproterozoic New Jersey Highlands along the New York–New Jersey border (Fig. 9). These granites resemble Lyon Mountain granite mineralogically, chemically, and in terms of their late- to postemplacement (Gorring et al., 2004). Multigrain zircon TIMS dating of the Mount Eve granite (Drake et al., 1991) indicates emplacement at 1020 ± 4 Ma (upper intercept on a three-point concordia plot). Within the New Jersey Highlands, undeformed anatectic meta-trondhjemites derived from the ca. 1.3 Ga Losee Suite have been dated by SHRIMP at 1030 ± 12 Ma, thus providing evidence for Ottawan thermal effects (Volkert et al., this volume).

Owens and Samson (2004) used Sm-Nd to produce an isochron age of 1046–1023 Ma for the widespread State Farm gneiss within the Goochland terrane to the east of the northern Blue Ridge (SFG, Fig. 9). Owens and Tucker (2003) reported a single-grain TIMS zircon age of 1057–1013 Ma for the same rocks. Aleinikoff et al. (1996) obtained a single-grain TIMS zircon age of 1045 ± 10 Ma for the Montpelier anorthosite that occurs within the State Farm gneiss, indicating that the association is coeval and probably represents a CRUML-type (Chateau-Richer, St. Urbain, Mattawa, and Labrieville) anorthosite-mangerite-charnockite-granite suite. The Maiden gneiss is widespread throughout the Goochland terrane, and had been considered to be Mesoproterozoic; however, Owens et al. (this volume) propose that most of it is Devonian in age. The Montpelier anorthosite shares geochemical affinities, including its Pb isotope characteristics (Sinha et al., 1996) with the ca. 1045 Ma Roseland anorthosite (Pettingill et al., 1984) of the northern Shenandoah massif (Owens and Samson, 2004).

Farther south, in the Smoky Mountains of the western Blue Ridge (Fig. 9), Tollo and Aleinikoff (this volume) report ages of 1045 ± 6 Ma, 1037 ± 5 Ma, and 1022 ± 4 Ma from weakly foliated late Ottawan intrusions. In the Mount Rogers area of southwest Virginia and northwest North Carolina (Fig. 9), Tollo et al. (2006) dated granitic intrusives at ca. 1060 Ma and noted that these silica-rich granitoids show evidence of only weak Proterozoic metamorphic effects. Within the Grandfather Mountain Window (Fig. 9), the Blowing Rock gneiss yielded a SIMS age of 1081 ± 14 (Carrigan et al., 2003). Investigations by Heatherington et al. (2006, 2007) and Steltenpohl et al. (2004) indicated that high-grade granitic gneiss of the Pine Mountain window (Fig. 3) yields ages of 1011 ± 12 Ma to 1063 ± 14 Ma. Ages of xenoliths within these rocks fall into two distinct groups: 1018 ± 24 Ma and 1136 ± 13 Ma. The younger granitoid ages are the youngest in the southern Appalachian Blue Ridge. Neodymium model ages indicate a range from 1.32 to 1.59 Ga for the granitoids and 1.35–1.97 Ga for the xenoliths; the latter suggest an exotic component (Heatherington et al., 2006). Hf model ages are 1430–940 Ma for the granitoids and 1.87–1.4 Ga for the xenoliths. These are similar to Mars Hill model ages and suggest that older, exotic crust may extend beyond there (Heatherington et al., 2007). Detrital zircons in the overlying Lower Cambrian Hollis quartzite also suggest exotic provenance, notably Amazonia (Steltenpohl et al., 2004).

Summary of Geochronology of the Mesoproterozoic Inliers of the Appalachians

When discussing the Mesoproterozoic inliers of the Appalachians, it is always necessary to keep in mind that these rocks were telescoped and shuffled by Paleozoic orogenesis so that their current distribution may bear little relationship to their original locations (Trupe et al., 2004). Notwithstanding this, they remain sufficiently coherent to provide significant regional correlations and offer the promise of even greater insights in the future. Most importantly, magmatism belonging to events in group 1 (1183–1144 Ma), group 2 (1143–1111 Ma), and group 3 (1078–1028 Ma) is basically the same from the French Broad massif to the Mount Holly complex, Vermont, and the adjacent Adirondack Mountains (Fig. 3). Magmatic group 2 is limited in extent and therefore difficult to correlate regionally. Throughout the Appalachians, groups 1 and 3 encompass deformation and metamorphism of the 1180–1140 Ma Shawinigan orogeny and the 1090–1020 Ma Ottawan orogenic phase of the Grenvillian orogeny, respectively (Tollo et al., this volume; Southworth et al., this volume). Local differences in timing and metamorphic grade exist, but the principal architecture remains solidly established. Similarly, the occurrence of ca. 1.4–1.3 Ga tonalitic, trondhjemitic, and granitic rocks has been well established from the Mount Holly complex to the New Jersey Highlands (Ratcliffe and Aleinikoff, 2008; Volkert et al., this volume), and a southern occurrence has now been reported from the Mount Rogers area of the French Broad massif (1327 ± 7 Ma; Tollo et al., this volume). All of these recently documented relationships provide an encouraging backdrop for future breakthroughs in understanding the geologic evolution of North America from Labrador to Alabama.

REGIONAL ISOTOPIC RELATIONS AND TECTONIC INTERPRETATIONS

A common feature of the Grenville orogen from Labrador to Alabama is the presence of metamorphic ages of ca. 1080–1020 Ma signifying Ottawan metamorphism in much of the belt.

There does not appear to be any systematic pattern of these ages along most of the orogen, and 1010–980 Ma Rigolet ages occur throughout the parautochthonous belt without preference to locality. This suggests that by late Ottawan time, eastern Laurentia had become a single cohesive orogenic belt. However, this was not always the case, and isotopic data discussed in the following suggest a complex evolution.

Neodymium Model Ages

Using drill-hole core studies, Bickford and Van Schmus (1986) and Van Schmus et al. (1989, 1993), demonstrated that the central North American basement consists largely of granite and rhyolite and may be divided into the Eastern granite-rhyolite province (EGRP, Fig. 11) and Southern granite-rhyolite province (SGRP, Fig. 11), which are separated on the basis of U-Pb crystallization ages: 1.4–1.5 Ga for the former and 1.37–1.4 Ga for the latter. A northeast-southwest line further divides the provinces (Fig. 11) and is based on Nd model ages. To the northwest of this line, Nd T_{DM} is >1.55 Ga, and to the southeast Nd T_{DM} is <1.55 Ga. Thus, in much of the Eastern granite-rhyolite province, the difference in time between crustal formation and magmatic intrusion was <100 m.y., and these plutons (e.g., in the St. Francois Mountains) are juvenile and related to a continental margin arc or backarc basin (Menuge, 2002; Rivers and Corrigan, 2000; Rohs and Van Schmus, 2007; Slagstad et al., 2009). In contrast, the 1.4–1.37 Ga plutons of the Southern granite-rhyolite province are much younger than the crustal formation age and represent reworked crust that may mark the eastern limit of the Paleoproterozoic interior (Van Schmus et al., 1993). The crust of the Eastern granite-rhyolite province is known to extend into Kentucky (Fig. 9), where a deep well yielded samples with U-Pb ages of 1381 ± 27 Ma and 1457 ± 10 Ma (Van Schmus et al., 1993). This locality is within 250 km of the western edge of the Western Blue Ridge Province. Given their proximity, the crustal (<1550 Ma) and plutonic ages (1400–1500 Ma) of the Eastern granite-rhyolite province might be expected to extend into the Mesoproterozoic inliers of the western Blue Ridge. However, this is not the case in either the southern or eastern Blue Ridge Provinces (Fig. 9). Within the former (including the Mars Hill terrane), T_{DM} values range from 1.82 to 1.34 Ga, and in the latter ages range from 1.88 to 1.5 Ga. Of 18 T_{DM} determinations in both terranes, 14 exceeded 1.55 Ga (Carrigan et al., 2003). Accordingly, unlike the Eastern granite rhyolite province; there appear to be few juvenile rocks in the southern Appalachians. In addition, neither the southern nor northern Blue Ridge terrane contains the granitic plutons of 1.5–1.4 Ga age that are so common and characteristic of the Eastern granite rhyolite province. Nd model ages for Roan Mountain in the Mars Hill terrane range from 1.65 to 2.3 Ga (Fig. 9) and are inferred to reflect an exotic origin (Carrigan et al., 2003; Ownby et al., 2004). As stressed by Tohver et al. (2004) and Loewy et al. (2003, 2004), the similarity in Nd isotope values between the French Broad massif (>1150 Ma) and Amazonia are striking, especially when contrasted with the pronounced difference of both of these with data from the Adirondacks, Llano, and midcontinent (Fig. 11).

Within the northern Shenandoah massif of the Virginia Blue Ridge, recent studies (Tollo et al., 2006; Southworth et al., this volume) have provided a spectrum of Nd T_{DM} model ages (Fig. 9) ranging from 1.62 to 1.37 Ga, with 9 out of 11 samples exceeding

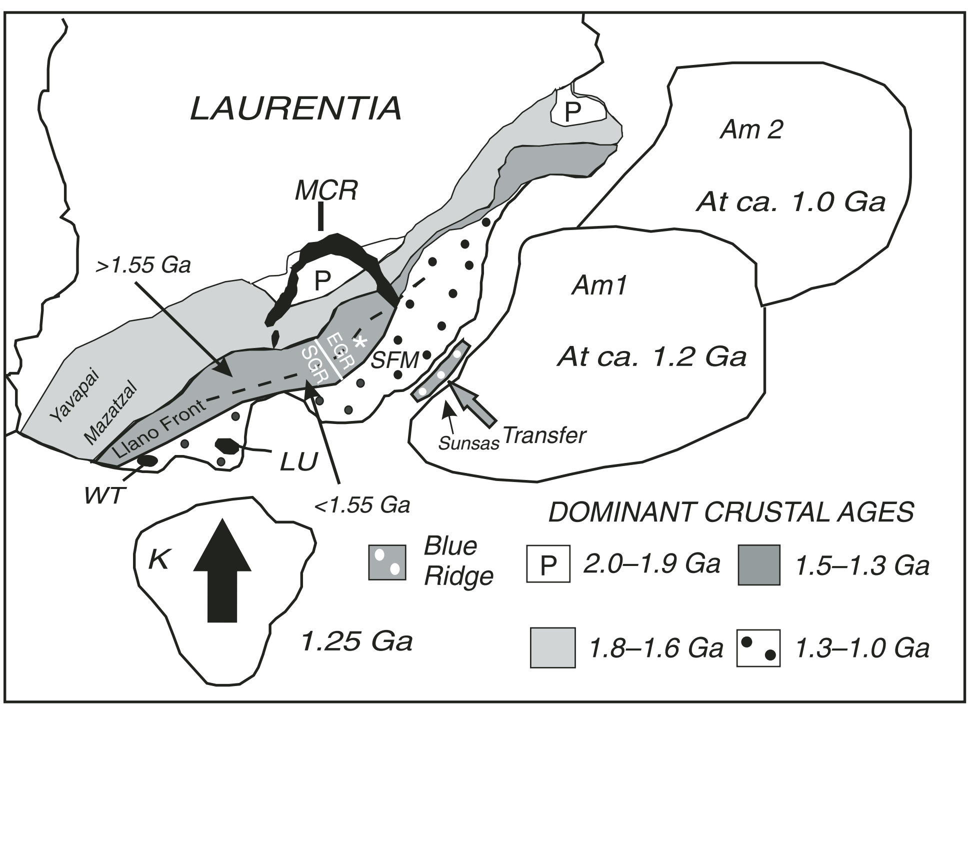

Figure 11. Schematic map of Laurentia showing some major crustal provinces and the proposed collision of the Kalahari craton (K) with the Llano uplift–west Texas area (LU, WT) at ca. 1.25 Ga and Amazonia (Am1 and Am2) with eastern Laurentia at ca. 1.2–1.0 Ga. The arrow pointing the Blue Ridge toward the southeastern Laurentian margin represents the transfer of Amazonian crust to Laurentia. The heavy dashed line separates regions with T_{DM} >1.55 Ma from those with T_{DM} <1.55 Ma as indicated. MCR—Midcontinent Rift; P—Penokean Province; SGR—Southern granite-rhyolite province; EGR—Eastern granite-rhyolite province; SFM—St. Francois Mountains (modified from Karlstrom et al., 2001; Loewy et al., 2003; Tohver et al., 2004).

1.49 Ga. The average of 21 Nd T_{DM} ages is 1.51 Ga. This age overlaps that of the Eastern granite rhyolite province, but, as noted, the area lacks the 1.5–1.4 Ga plutons of the former and hence must have been east of the continental margin arc. Thus, it seems unlikely that these ages represent an extension of the Eastern granite-rhyolite province. In general, Nd crustal ages in the Shenandoah massif are older than those of the Adirondack–Mount Holly crust and suggest a different origin. However, the Nd model age evidence is insufficient to categorize the Shenandoah massif as exotic, especially since the southern half of the massif has not received detailed attention. Nevertheless, the possibility of an exotic origin is strengthened by whole-rock Pb isotopic data discussed later herein.

Pb Isotope Arrays, Basement Mapping, and the Amazonian Collision

Sinha et al. (1996) and Sinha and McLelland (1999) undertook whole-rock Pb isotopic studies of the southern Appalachians and the Adirondack mountains, respectively. The results of the two Pb isotope investigations are shown in a ($^{206}Pb/^{204}Pb$) versus ($^{207}Pb/^{204}Pb$) plot (Fig. 10), and they indicate that the Adirondacks and the Appalachian Blue Ridge Province are not located above, and were not formed from, a single, commonly shared Pb reservoir. Data from the Adirondacks plot below the continental Stacey-Kramers continental growth curve (Sinha and McLelland, 1999), whereas data from the Blue Ridge and Amazonian analyses overlap and plot mostly above the Stacey-Kramers curve (Sinha et al., 1996). The Honey Brook uplift in Pennsylvania (Fig. 10) plots below the curve and very close to the Adirondack samples. Lack of overlap indicates that the two regions have different ancestry (Tohver et al., 2004). The Adirondack data coincide with results from the Eastern granite rhyolite belt and the Llano uplift in Texas (Roller, 2004). Loewy et al. (2003) obtained 44 whole-rock Pb isotope ratios from Labrador, and these demonstrate that values similar to those observed for the Adirondack values continue to the northeastern edge of the Grenville Province. The similarity among the Pb isotope plots of the Adirondacks, Grenville Province, Llano uplift, and Eastern granite-rhyolite province are as striking as their differences from the Blue Ridge and Amazonia. These similarities, as well as those in the Nd model ages, have resulted in the proposal that much, and perhaps all, of the Blue Ridge was transferred from Amazonia (Fig. 11) to eastern Laurentia during a ca. 1.2 Ga collision (Loewy et al., 2003, 2004; Tohver et al., 2004, 2006, and references therein).

Hoffman (1989) proposed that the ca. 1090–1030 Ma Ottawan orogeny resulted from a head-on Himalayan-style collision between Laurentia and Amazonia, the Precambrian core of Brazil. This model was largely based on downdip lineations on the Grenville Front, paleomagnetic data, and similarities in age and lithologic character between the Grenville Province and the Sunsas, Aguapie, and Nova Brasilandia belts of the Amazon craton (i.e., Amazonia), where 1.2–1.0 Ga metamorphism is recognized (Tohver et al., 2002, 2004). However, the model is apparently inconsistent with paleomagnetic studies on flat-lying, ca. 1.2 Ga gabbros and basalts of the Nova Floresta Formation in western Brazil (Tohver et al., 2002). This investigation yielded a pole that falls on the combined 1.3–1.15 Ga apparent polar wander path (APWP) for Laurentia and Greenland at ca.1.2 Ga and places Amazonia against the Llano uplift, suggesting an early Grenville collision between the two terranes (Tohver et al., 2002, 2004). More recently, D'Agrella-Filho et al. (2007) calculated a new paleomagnetic pole for red beds in the Aguapei Group, from which xenotime rims on detrital zircons provide a well-constrained age of 1149 ± 7 Ma for diagenesis. This result places the southwest Amazon craton near the southeast Appalachian portion of Laurentia at ca. 1150 Ma, and appears to be the most compatible with the Pb and Nd relationships discussed already.

The regionally anomalous Nd T_{DM}, zircon crystallization ages, and Pb-Pb whole-rock data in the Mars Hill terrane are distinctive in eastern Laurentia (Ownby et al., 2004; Geraldes et al., 2001) but are similar to features in the 1300–1000 Ma Sunsas belt of southwestern Amazonia (Loewy et al., 2003). The similarities are striking and support the Amazonian derivation of the Blue Ridge–Mars Hill terrane. Given this, the "major ca.1.2–1.18 Ga disturbance" documented by Ownby et al. (2004) in the Mars Hill terrane likely records the initial collision of Amazonia and Laurentia, and the onset of transfer of Amazonian crust to eastern North America. The Amazonian-Laurentian collision from ca. 1.2 Ga to 0.98 Ga took place as the Amazon craton underwent sinistral, transpressional collision characterized by pulses in local tectonic activity as the craton progressed along a northeastward path bordering Laurentia (Fig. 11).

Loewy et al. (2003) proposed that the Kalahari craton represents the "southern continent" that Mosher (1998) and Mosher et al. (2008) suggested collided with the Llano uplift at ca. 1.25 Ga (Fig. 11). Shortly thereafter, at ca. 1.2 Ga, Amazonia underwent a sinistral transpressional collision with southeastern Laurentia, and the transfer of terranes to the Blue Ridge took place.

Arguments for continental collision between Amazonia and Laurentia with transfer of crust to the latter are compelling and consistent with a wide range of evidence. However, important details remain unclear and require resolution. These include delineation of the extent of Amazonian crust as well as the degree to which it was reworked to produce Laurentian magmatism now represented by 1.2–1.0 Ga zircon ages. Solutions to the problem will require more surface mapping and geochronology, including Nd and Pb isotopic studies. Nevertheless, the existing and continually evolving geologic database is beginning to provide insight into the Mesoproterozoic history of the Appalachian inliers and make regional correlation of events and processes possible.

CONCLUSIONS

The Grenville Province evolved over an extended period (ca. 2.0–1.3 Ga) that included long-lived continental margin arcs located along the southeast margin of Laurentia. By 1.3 Ga, rifting produced the Central metasedimentary belt backarc basin and associated island arcs (Fig. 7). At ca. 1.1 Ga, collision of Amazonia with this region resulted in the Grenvillian orogeny and development of a large, hot, long-lived orogen similar to the Himalayan orogen. Culminating Ottawan-phase contraction took place from 1090 to 1050 Ma, when large allochthons were emplaced toward the foreland along the allochthon boundary thrust. By 1020 Ma, the hot middle crust had become so ductile that the hinterland underwent extensional plateau collapse. Ensuing Rigolet-phase contraction at ca. 1010 Ma gave way to further collapse at 980 Ma and resulted in formation of the Grenville Front thrust system. Grenvillian effects are represented from Labrador to Alabama and Texas.

The Adirondack Highlands and inliers of the northeastern Appalachians from the Mount Holly complex, Vermont, to at least as far south as the southern New Jersey Highlands contain ca. 1.4–1.3 Ga tonalitic to granitic orthogneiss representing fragments rifted from the Laurentian continental-margin arc at ca.1.3 Ga during the inception of the Central metasedimentary belt backarc basin (Figs. 7A). We propose that the approximately north-south backarc basin represents the western arm of a larger approximately east-west rift (Gower, 1996) that separated Adirondis from eastern Laurentia by 1.3 Ga (Figs. 7A and 7B). East-dipping subduction related to basin closure took place beneath the Mount Holly complex to the New Jersey Highlands (Fig. 7B) and led to ca. 1270–1250 Ma Elzevirian calc-alkaline magmatism in these regions. At ca. 1220 Ma (Fig. 7C), accretion of the Central metasedimentary belt to Laurentia resulted in a stepping outward of subduction to the east, with a westerly polarity beneath the Adirondack Lowlands (Wasteneys et al., 1999; Peck et al., 2004). Chiarenzelli et al. (2007) reported a large volume of ophiolitic assemblages including oceanic peridotite together with sulfide deposits and thin cherty quartzites within the Adirondack Lowlands (X in Fig. 4), which is consistent with this model (Fig. 7C). Metamorphic zircons from the peridotite yield an age of 1202 ± 10 Ma, providing a minimum age that is significantly younger than the surrounding metapelites, which have depositional ages of <1300 Ma based on detrital zircons (Chiarenzelli et al., 2007). Thus, an age of ca. 1.3–1.2 Ga for the ophiolite appears to be reasonable and is consistent with the existence of an ensimatic arc at that time. If the Honey Brook anorthosite (Fig. 9) is shown to be ca. 1155 Ma in age, then the arc may have extended to at least as far south as the Baltimore gneiss domes. In that case, the dividing line between Laurentian and possibly Amazonian crust would be located south of the Honey Brook uplift and then track to the NE parallel to the Grenville Province (dashed line, Fig. 9). If the Mount Rogers 1327 ± 7 Ma granofels is shown to be part of the Dysart–Mount Holly Belt, then an even more expansive model will be necessary.

The earliest evidence of Shawinigan deformation and magmatism occurs at ca. 1.2 Ga within the Mars Hill terrane and is inferred to have coincided with the initial collision of Amazonia with southeast Laurentia (Ownby, et al., 2004). This age corresponds to the onset of pre-Shawinigan magmatism in the Adirondack Lowlands during closure of the Central metasedimentary belt backarc basin (Fig. 7C). We suggest that both of these events are manifestations of the Amazonian collision. We speculate that the ca. 1.2 Ga collision of Amazonia and Laurentia took place approximately normal to their north-northeast–trending, subparallel margins (Fig. 11) and resulted in intense orogenesis throughout the Appalachians and Adirondacks, including the short, western extension of Adirondis that contains rifted fragments of the ca. 1.4–1.3 Ga Laurentian margin arc. In contrast, the long, eastern arm of Adirondis would have moved subparallel to the Laurentian margin (Figs. 7C and 7D), and, therefore, that region would have experienced less deformation and magmatism. In the case of the Grenvillian orogeny, the head-on collision would have been along the approximately east-west–trending Laurentian margin of what is now the Grenville Province (Fig. 7F), and the most intense deformation and metamorphism would have occurred in that region. In much of the Appalachians, the subparallel northeast-trending margins of Laurentia and Amazonia (Fig. 11) would have resulted in relatively lower intensity orogenesis during northeasterly relative displacement, consistent with the reduced level of ca. 1070–1050 dynamic metamorphism reported in the Shenandoah massif (Southworth et al., this volume).

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